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The Eastern Bering Sea Shelf: Oceanography and Resources

Volume One

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The Eastern Bering Sea Shelf

The Eastern Bering Sea Shelf: Oceanography and Resources

Edited by Donald W. Hood and John A. Calder

Volume One

UNITED STATES
DEPARTMENT OF COMMERCE
Philip M. Klutznick, Secretary

NATIONAL OCEANIC AND
ATMOSPHERIC ADMINISTRATION
Richard A. Frank, Administrator

*OFFICE OF MARINE POLLUTION
ASSESSMENT
R. L. Swanson, Director*

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Foreword

President Nixon, in his energy message of 23 January 1974, set up Project Independence, the goal of which was to make the nation self-sufficient in oil production by the end of the 80's. Oil and gas deposits on the outer continental shelf of the nation were estimated to be the largest single source of petroleum available. Consequently, the expeditious development of the petroleum resources of the outer continental shelf became an essential and central element in the plan to attain the goal of Project Independence. Because of the extensive outer continental shelf of Alaska and the high estimated potential petroleum resources contained therein, Alaska was quickly placed at the forefront of the nation's OCS oil and gas development program. Potential lease areas or basins were identified encompassing significant parts of Alaska's vast OCS area extending from the northeast Gulf of Alaska to the demarkation line in the Arctic.

Several of the basins of most promise are in the Bering Sea, an area of exceptionally high biological productivity which historically has provided bountiful harvests of fish and other living resources to the United States and other nations. The Bering Sea supports a wide variety of marine mammals and birds, many legally designated as endangered species, and others harvested commercially by foreign nations. Alaskan natives rely heavily on many of the Bering Sea living resources as the cornerstone of their subsistence culture.

The requirement to develop sufficient energy sources for the present and future decades has been identified as certainly one of the paramount problems of expanding urgency which our nation faces. Development of the petroleum resources of the Bering Sea Outer Continental Shelf offers a promise of substantial benefits to the nation. Associated with these developments is an element of risk to the marine environment and living resources of the area. Public concern for the conservation and protection of the marine environment and its living resources, especially in the pristine environments of the Alaska OCS area, is well recognized. Decisionmakers in the private sector as well as in government must be responsive to the expressed concern about possible environmental and resource damage caused by OCS oil and gas development. Decisionmakers are required by law to develop an understanding of the possible risks to the environment and resources, to quantify, where possible, the probability and extent of likely damage, and to utilize this information and understanding in their decisions about OCS oil and gas development.

As manager of the Outer Continental Shelf Leasing Program, the Bureau of Land Management (BLM) of the Department of the Interior (DOI) initiated in 1975 the Alaska Outer Continental Shelf Environmental Assessment Program (OCSEAP). OCSEAP is a comprehensive, multidisciplinary environmental studies program designed to provide BLM in particular, other decisionmakers, and the interested public with a source of adequate information to help them formulate their decisions and to generate management strategies that provide acceptable protective and mitigating measures against the possible ranges of undesirable or unacceptable impacts on the marine environment and living resources. OCSEAP is managed for BLM by the National Oceanic and Atmospheric Administration (NOAA) through an interagency agreement.

An important goal of the environmental studies program in Alaska is to enable the BLM to organize and develop policy guidance for major decision points in their OCS leasing process. The implied premise is that sound policy governing OCS leasing and petroleum development will result from adequate knowledge and understanding of the environment, the ecological processes controlling the distribution and relative abundance of important populations of marine organisms, and their vulnerability and sensitivity to OCS petroleum development. In the management of OCS oil and gas development it is recognized that the viability of policies will be enhanced if both what is known and what is not known are included in the decisionmaking process. It also is apparent that the most effective and economical use of OCS study funds can be achieved when there is a clear understanding of the status of scientific knowledge in the region of concern. This also includes both what is known and what is not known. If the status of knowledge is sufficiently documented, it enables the studies managers to identify information needs that are attainable within limits of resources and time available, and which have the greatest relevancy to the issues of concern associated with OCS oil and gas development. The rationale above was the primary consideration which caused OCSEAP/BLM to produce the treatise, *The Eastern Bering Sea Shelf: Oceanography and Resources*.

As recently as 1970, the Bering Sea was largely a frontier area for marine science. At that time most of the research in the Bering Sea had been conducted by the Russians, with lesser contributions by Japan and the United States. Emphasis was primarily on the assessment of commercial species of fish and shellfish with lesser efforts directed toward other marine organisms, such as marine birds and mammals, biological productivity, and the physical environment. In 1974, Hood and Kelley published a review of knowledge existing before 1970, *Oceanography of the Bering Sea with emphasis on renewable resources*.

Within the last decade our knowledge and understanding of the Bering Sea have been significantly expanded. The BLM/NOAA Outer Continental Shelf Environmental Assessment Program contributed substantially to our increased understanding of the outer continental shelf in the Bering Sea and the living marine resources of the area.

The present publication consists of a selection of 73 chapters written by acknowledged experts in their fields, giving the most accurate and comprehensive description to date of the physical environment and resources of the outer continental shelf of the Bering Sea from the Aleutian Islands to the Bering Strait. The book should be invaluable as a source of scientific data and information readily available for use in decisions regarding the management of OCS petroleum development. It will be equally valuable to OCSEAP in planning and conducting continuing environmental studies in support of BLM's needs for environmental and resource data to guide policy decisions. Finally, the book will be of great interest and use to the scientific community in conducting a wide range of research projects in future years.

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Alaska Outer Continental Shelf Office

Herbert E. Bruce
Director, Alaska Office, Office of Marine Pollution Assessment,
National Oceanographic and Atmospheric Administration

Preface

The production of these volumes, entitled *The Eastern Bering Sea Shelf: Oceanography and Resources*, was conceived by the Juneau Project Office of the Outer Continental Shelf Environmental Assessment Program (OCSEAP) with the aim of providing a compilation of basic information on the Bering Sea Shelf to be used by the Bureau of Land Management in its primary responsibility for implementing most of the pre-sale oil and gas lease objectives. Of even greater long-term value for the protection of the Bering Sea environment, this treatise will help to build an understanding of how the Bering Sea functions as a system. Only by understanding how this ocean functions will we be able to predict the impact of human activities on specific parts or the whole of this unusually productive environment.

It is the primary purpose of this publication to present in a single document what is now known about the natural science of the Eastern Bering Sea Shelf. Some criticism has arisen that this effort is premature—that so much work is presently in progress that a later treatment would be more definitive. It might be desirable to wait until present work is finished, reflected upon, and analyzed, but it is not the nature of interdisciplinary scientific effort for all disciplines to reach a comfortable plateau in status of knowledge simultaneously, allowing for integration, substantive analysis, and thorough documentation. Instead, one discipline feeds on the information of another as it becomes available and understandable, thus building toward an understanding of the way the system functions as a whole. Apart from the dictates of the workings of interscience disciplines, the expediency of the times demands that the best information possible be available for use in formulating impact statements for oil and gas development in the Bering Sea during 1980 and beyond.

We have, therefore, done the best we could to set down all we know about the natural science of the Bering Sea Shelf. It is apparent that our scientific knowledge of this region is well advanced and that the analysis of certain of its elements is at a level equal to or exceeding that of any like region in the world. Our efforts here have not produced the ultimate document of the Bering Sea Shelf—one may never be written—but this is intended to be an important benchmark in Bering Sea literature upon which future science and the use of science for pragmatic purposes can be based.

A secondary, but important, purpose of these volumes is to provide a credible scientific document from which estimates of the effects of oil and gas development in the outer continental shelf region of the eastern Bering Sea can be made. The creation of this work has provided a medium for presenting basic scientific evidence about how the natural system works, uncomplicated by the requirement of applying the findings to specific problems of oil and gas development. The application of this information is being accomplished through other workshop and synthesis efforts.

Considerable time and expense are required to produce a high-quality publication containing the essential information, and such a project must usually be supported by a mission-oriented department of the federal government. Despite what may be administrative difficulty, it is important that provisions be made to allow for appropriate scientific contribution to decisionmaking processes involving the uses of the ocean, whether it be for oil and gas development or for other human needs.

The mechanism for preparing this document consisted of several steps. First, the extent of the pertinent information available on the Bering Sea Shelf had to be determined. Knowledgeable Bering Sea scientists were assembled; they explored the amount of material that could be brought together for publication at this time, and agreed to serve as associate editors of thirteen disciplinary and interdisciplinary sections. This gave us a tentative outline. Next these same scientists met again to propose authors and extent of material in their respective areas. A symposium was held in Anchorage, Alaska, in November 1979, on which occasion much of the material was presented orally, and some drafts of papers were collected which would eventually constitute some of the chapters presented here. By 15 January 1980 a draft of the total volume, consisting of fourteen sections containing sixty of the chapters in these volumes, was furnished to the Bureau of Land Management as an interim working document. Then began the long process of editorial, typesetting, composition, and printing effort to produce the final form of the book.

I hope that through this exercise we have all gained insight into making scientific information serve the environmental decision-making process so that all men, present and future, may properly benefit from the bounty of the ocean's resources.

Many people contributed to the organization and preparation of these volumes, but we list only those who have taken a leadership role in specific aspects of its production.

Printing preparation was done by Science Applications, Inc., in Boulder, Colorado, with Joseph G. Strauch, Jr., as principal investigator, Patricia Martin Gibby as technical editor, and Robert E. Peterson as graphics editor.

Much of the material included in this book was presented at a symposium held in Anchorage, Alaska, in November, 1979, organized by Leland Hepworth and John A. Calder, Jr., OCSEAP program managers for the Bering Sea Studies. Working groups at this symposium were led by E. Carmack for physical oceanography, P. Becker for benthic biology, D. Hood for chemical oceanography, D. Menzel for plankton ecology, B. Farentinos for birds and mammals, P. Fischer for geologic and ice hazards, D. Nyquist for ice-edge ecosystems, and O. Mathisen for fisheries biology.

Each of the fourteen sections of these volumes had an associate editor, named at the beginning of each section, responsible for selecting authors, comprehensively covering the subject, and organizing the section internally in such a way as to be coherent with the whole treatise.

Funding came from the Outer Continental Shelf Environmental Assessment Program, which is administered by the National Oceanic and Atmospheric Administration for the Bureau of Land Management. The encouragement and administrative support of H. E. Bruce, Juneau Project Manager, and R. L. Swanson, Director of the Office of Marine Pollution Assessment of NOAA for this effort are greatly appreciated.

D. W. Hood

J. A. Calder

Introduction

Donald W. Hood

In the eighteenth century, Kamchatka and Alaska were better known and much more the object of international attention than was California. Should wealth be the criterion?

William R. Hunt (1975)

Sandwiched between the northernmost land masses of the North American and Asian continents lies the Bering Sea. It has a mean depth of 1,636 m, a surface area of 2.3×10^6 km² and a volume of 3.7×10^6 km³. These dimensions make it the third largest semi-enclosed sea of the world ocean, surpassed in size only by the Mediterranean and the South China seas. It derives its name from Vitus Bering, who in 1725-43 commanded an extensive Russian expedition which explored the coasts of the Kamchatka Peninsula to the south of the Aleutian Islands and the southern mainland of Alaska as far as Prince William Sound (Hood and Kelley, 1974).

From north to south the Bering Sea can be viewed as a sector with a radius of about 1,500 km with the Bering Strait approximately at the vortex. The southern arc represented by the Aleutian Islands and the Alaska Peninsula extends between 157° W on the shores of Bristol Bay on the east to 163° E at the coast of the Kamchatka Peninsula on the west, a total distance of almost 3,000 km.

Included within the Bering Sea boundaries are three major bays: Bristol Bay in the southeast and Norton Sound in the northwest, both bordering Alaska; and the Gulf of Anadyr in the northwest, bordered by Siberia. Three major rivers—the Kuskokwim and Yukon draining central Alaska and the Anadyr draining western Siberia—empty into the Bering Sea. All these rivers contain glacial melt water and runoff from taiga forests and tundra characteristic of the far northern environment.

The Bering Sea is often considered the northern extension of the Pacific Ocean. It is true that exchange of surface water through the Aleutian Island passes occurs relatively freely and water characteristic of the north Pacific Ocean is evident within the Bering Sea. However, since water passage is limited by the sill depths of the passes except in the west and the link with the Arctic Ocean is the narrow (85 km), shallow (45 m) Bering Strait, this sea possesses the essential features of a well-defined ocean region and may be best looked upon as a con-

finer sea with characteristics of its own sufficiently different from adjoining oceans to make it a unique ocean region.

The eastern Bering Sea shelf (Fig. 1), about one-half (1.2×10^6 km²) the total area of the Bering Sea, is exceeded in size only by the shelf common to the Chukchi, East Siberian, and Laptev seas, within the geographic boundary of the Arctic Ocean. Besides its size, it has several other features which may contribute significantly to the way this shelf functions oceanographically to make it one of the most productive of the world's ecosystems.

Three features clearly differentiate the eastern Bering Sea shelf from most other shelves for which there is a sufficient data base for comparison. First, there is a recurring, although variable in extent, annual ice cover that in cold years reaches as far south as the Alaska Peninsula. Second, it is vented at the northern end by the Bering Strait, through which passes about 2×10^6 m³ of water per second (2 Sv). This water has its origin in the North Pacific, with a small contribution (10 percent) from Alaskan rivers, and yet little of it penetrates the southern shelf area. Third, there are three persistent fronts, most dominant in the summer season, that lie roughly over the shelf break (150 m) and the 100-m and 50-m contours in the southeast shelf region. These fronts have great influence on the southeast Bering Sea and possibly the entire shelf ecosystem. Other features such as topography, wind patterns, and general climate also contribute heavily to the function of this shelf.

The Bering Sea ice cover and ice dynamics are the subject of chapters 9-14 of this volume; these represent a substantial contribution to knowledge about physical and meteorological features that influence the dynamics of the ice system. When ice forms in the fall, it acts as a physical barrier to direct sea-air interaction and also as an effective insulator against energy transfer between the near-freezing shallow waters of the shelf and the overlying Arctic air mass. The formation of ice and the resulting increase in density of the surface water combine with wind mixing on the surface and tidal mixing at depth to cause complete vertical homogenization of the water column over the whole of the shelf during the early winter. The advances and retreats of the ice are variable, depending on broad-scale meteorological events, sea and air temperature, and surface winds. The period 1973-78 was a time of extreme fluctuations in ice conditions: the years 1973-76 were characterized by below-normal temperatures which brought about abnormally high ice coverage, which reached as far as Unimak Pass in 1976 and persisted well into May before sufficient warming occurred to

Figure 1. The eastern Bering Sea shelf.

bring on the spring melt. On the other hand, 1977-79 was a period of high air and sea temperatures and limited ice coverage brought on by southerly shifting winds (See Niebauer, Chapter 9, this volume). These widely varying physical conditions greatly influenced the primary productivity, and thereby all other biotic components, of the shelf region, as described in Sections VII and X of Volume 2.

The oceanographic significance of the shallow northern opening of the Bering Sea through the Bering Strait into the Chukchi Sea is yet to be fully

ascertained, particularly as it contributes to the whole of the biological system. The pathway to the Arctic from the Pacific through the Bering Sea has long been known to be of major importance to some migrating whales (See Frost and Lowry, Chapter 50, Volume 2). Furthermore, the significance of the Bering Strait passage for walrus, seals, and beluga and bowhead whales between the Arctic and Bering Sea is discussed by Burns, Chapter 46 and Lowry and Frost, Chapter 49 of Volume 2. Although some of the highest short-term primary production rates

ever measured in the world's oceans were found in the Bering Strait (McRoy et al. 1972), the processes which caused this production and its quantitative extent have not been elucidated.

The influx of water into the Bering Sea through the Aleutian passes has been summarized by Favorite (1974). Several conflicting opinions persist as to the amount and location of flow through the passes. Arsenev (1967) believed that the Komandorski/Near islands Strait is the most significant source of inflow (14.4 Sv) with a considerably reduced flow in the central Aleutian passes (4.4 Sv); no net exchange was considered to occur in the eastern Aleutian passes. It is clear that much of this water must return to the Pacific through deep passes (Hughes et al. 1974), since the outflow to the Arctic Ocean is only 1-2 Sv and the Bering Sea is generally considered to have a positive net influx of fresh water.

Of equal or perhaps greater importance to the productivity potential of the Bering Sea is the vertical transport (upwelling) caused by water flow through the passes as described by Hood and Kelley (1976) and Swift and Aagaard (1976). This transport was shown to be responsible for providing about 5×10^6 g of nitrate nitrogen per km^2/day to the surface waters of a region of several thousand square kilometers observed north of Samolga Pass. Other passes are expected to contribute similarly. The fate of nitrate and other accompanying nutrients brought to the surface by upwelling in the Aleutian passes has not been determined, but this upwelling could be another important reason for the high bioproductivity observed in the Bering Sea.

The third unique feature of the shelf, that of three ocean fronts, is a recent discovery made possible by detailed hydrographic surveys of the region through the Outer Continental Shelf Environmental Assessment Program (OCSEAP), sponsored by the Department of the Interior, and Processes and Resources of the Bering Sea (PROBES) studies sponsored by the National Science Foundation. The details of these oceanic structures are described in Kinder and Schumacher (Chapter 4, this volume) and the biological significance to plankton ecology by Goering and Iverson (Chapter 56, Volume 2) and Cooney (Chapter 57, Volume 2).

Historical setting

The north Pacific Ocean south of 49° N latitude between the coasts of America and Japan was fairly reliably described on maps dating from the late sixteenth century; however, almost total ignorance

prevailed about the region north of that transect. It was not until 1643 that a Dutch East India company expedition commanded by Maerten Gerritszoon Vries discovered a previously much talked about mystery land, Yezo, to the north of Japan—Yezo is now known as the Kurile Islands. In 1639 a bargeload of Cossacks made their way down a Siberian river to the Sea of Okhotsk for a first view of the "Icy Sea" to the north; the Amur, Ud, Anadyr, and Okhota rivers flowed south to the "Eastern Sea." Still there was no knowledge of a passage between the two bodies of water.

The question of whether America and Asia were united attracted the attention of Peter I, Czar of Russia from 1682 to 1725, who initiated a program of westernization by utilizing Swedish prisoners of war in Siberian ports as teachers of shipbuilding and navigational arts. After the death of Peter I, the program was continued by his successor, Empress Catherine, when Vitus Bering, a Danish explorer, and Lt. Alexei Chirikov were commissioned for their first expedition. In 1728, Vitus Bering, with the two ships *St. Peter* and *St. Paul*, sailed through the Bering Strait to settle once and for all that Asia and America were separated. On the second voyage, fatal to Vitus Bering, who died of scurvy in 1841, much of the Arctic coast of Asia was charted, as well as the southwest Alaskan coast and the Aleutian Island arc. Even though Vitus Bering had remarkable success, his work was doubted and in fact discarded to some extent. The pressing question of a northern passage from the Pacific Ocean led the British Admiralty to issue instructions to their most talented explorer, Captain James Cook, to extend his third voyage in 1778-80 in further search of a northeast or northwest passage, from the Pacific into the Atlantic Ocean or the North Sea of Europe by way of Russia. At the helm of the *Resolution* and accompanied by Lt. Charles Clarke on the *Discovery*, Cook passed through the Bering Strait into the Arctic Ocean in the summer of 1778. The expedition penetrated to 70° N, where Cook viewed the "Chuckchee" Sea and the extremities of both continents, and the possibility of a northwest or northeast passage to India so long sought from the Atlantic Ocean side had to be renounced. The remarkable accuracy of Bering's earlier cartography was demonstrated. The expedition, on its way south to wait out the Arctic winter, visited Norton Sound and many inlets in south-central Alaska. The following summer's expedition again sailed through the Bering Strait, but without Captain Cook, who had been murdered during the winter of 1779 by Sandwich Island natives.

Physical dimensions and ice cover

The Bering Sea is divided into a neritic area (0-200 m) and an abyssal region (over 1,000 m) of about equal extent, the sum totaling 87 percent of the entire area of 2.3×10^6 km². The continental slope, representing only 13 percent, generally has a slope of 4-5° and essentially divides the Bering Sea in half in a northwest-southeast direction. The shelf feature of the eastern half continues north through the narrow Bering Strait into the Chukchi Sea. The shelf region contains several islands, important to marine mammal and bird ecology, which influence the circulation of water and the formation and movement of sea ice. St. Lawrence, the largest of the five major shelf islands, is nearly 200 km long and lies between Norton Sound and the Gulf of Anadyr, south of the Bering Strait. Nunivak Island, the second largest, lies near the coast of Alaska between the mouths of the Kuskokwim and Yukon rivers. St. Matthew, St. Paul, and St. George islands, the other three large islands, lie well offshore. St. Matthew is about midshelf, approximately 275 km southwest of St. Lawrence, whereas St. Paul and St. George, the fur seal islands, are near the shelf break some 300 km northeast of Unimak Pass. The deep basin of the Bering Sea, beginning west of Unimak Pass, is separated from the Pacific Ocean by the Aleutian Island arc, which rises from 7,600 m on the Aleutian Trench side and 4,000 m on the Bering Sea side. Passes through the arc deepen westward, the deepest being 4,420 m between Kamchatka and the westernmost Komandorski Island. North of this passage a sill at 3,589 m provides the deepest passage to the basin. Except for a narrow passage at 180° E, the sill depth east of 171° E is much less than 1,000 m. The total cross section permitting exchange of water through the southern boundary is only 701 km² (Favorite 1974).

The Bering Sea basin is a vast plain lying at a depth of 3,800-3,900 m with occasional gradual sloping hollows to depths of as much as 4,151 m. Two submarine ridges penetrate the basin. The Shirshov Ridge extends south from Kamchatka along 170° E longitude to near the Aleutian arc and separates the eastern and western Aleutian Basin. The North Rat Island Ridge (Bowers Bank) extends 300 km north in a counterclockwise direction.

The four major rivers discharging into the Bering Sea are the Yukon, Kuskokwim, Kamchatka, and Anadyr. The flow of these rivers is much greater in the summer months, since the major runoff in all these drainage basins is from melt water. The Yukon, the largest, has a peak flow in August about equal to that of the Mississippi, and a mean flow for the

year of about two-thirds that of the Columbia (for details see Ingraham, Chapter 29, this volume).

Nonrenewable resources

The nonrenewable resources most likely to be found in the Bering Sea are hydrocarbons and heavy metals. About 75 percent of the shelf area holds some likelihood of hydrocarbon accumulation, and some areas of the shelf are extremely promising. The most promising areas are the St. George basin on the outer margin of the continental shelf, the North Aleutian basin, the Navarin basin, and the Norton basin. These sites have a long geological history of high organic productivity. The outer basins have 1-2 km of Oligocene to Recent deposits, mainly consisting of diatomite and diatomaceous sand, with some conglomerate. The North Aleutian Basin contains several kilometers of terrigenous and volcanogenic sediments. The Norton Basin was probably inundated during the late Miocene, and since the basin was probably receiving continental sediments from the late Oligocene, reservoir rocks should be common. The Yukon River has been a major source of sediment since the middle Miocene, and probably large sources of natural gas are present.

The occurrence of gold, platinum, and tin placers on the shores of the Bering Sea is well known, but exploration for offshore deposits has thus far been disappointing. Almost all placers found have been in the nearshore zone, generally less than 5 km from shore. The only significant discoveries are rich, but small, submerged deposits of placer gold off Bluff and much larger, low-grade gold deposits on the sea floor off Nome. Commercial production from the latter deposits began in 1975. The search for tin in and near the Bering Strait and for platinum in and near Goodnews Bay has yet to yield more than minor results.

Renewable resources

The borders and islands of the Bering Sea have been inhabited by Eskimos in the northern parts and by Aleuts along the Aleutian arc as long as man has a historic record. These natives subsisted on a hunting and fishing economy for centuries in compatibility with an abundance of fish and animals in the area. The ancient cultures are still viable in some areas (Fig. 2). Not until western man began exploiting the region in order to supply large populations of people with furs, fish, and marine mammals was the natural living wealth of the area affected. As a result the most valuable asset, the sea otter, was exploited to the point of near extinction (Schneider, Chapter 51, Volume 2), and the Steller sea cow lasted

Figure 2. Eskimos walrus hunting in the Bering Sea (photograph courtesy of John J. Burns).

only 27 years between discovery and extinction (Hunt 1975). At least part of the herring's ecological niche appears to have been replaced by the Alaskan pollock (Wespestad, Chapter 32, this volume); the sockeye salmon (Straty, Chapter 35, this volume), once endangered and still under heavy pressure, is being restored, but international cooperation will be necessary in order to avoid further, even accidental, depletion of the stocks. Throughout recent history the wide diversity of the living resources of this region has been known, but only recently has it been fully appreciated (Favorite, Chapter 28, this volume). Large harvests of pollock, cod, ocean perch, black cod, halibut, rattails, tanner and king crab, and, more recently, shrimp are now being taken by many of the major fishery nations of the world. Until

recently Japan harvested by far the most, followed by the U.S.S.R.; but since the passage of the 1977 Fisheries Conservation Act the United States is catching up and now clearly dominates the crab fishery. The commercial catch of finfish is discussed by Bakkala (Chapter 61, Volume 2) and of crabs by Otto (Chapter 62, Volume 2). The potential for other fisheries is treated by McDonald et al. (bivalve mollusks, Chapter 67); Hughes (surf clams, Chapter 68) and MacIntosh and Sowerton (gastropods, Chapter 69, Volume 2). The large annual tonnage constituting the human harvest of fish represents less than 4 percent of the nekton and benthos required to support the extremely large population of mammals and birds dependent upon this resource (Laevastu and Favorite (Chapter 37, this volume).

Conclusion

Due to its latitude, the Bering Sea lies in a region with great annual variations in properties. Incident radiation in the northern region varies annually from almost total daily darkness to total light; the wind torque over the sea is an order of magnitude greater in winter than in summer; and the extensive ice cover in the winter is totally absent in the summer. Although warm seas contain more diverse populations, the colder seas support much larger individual populations. The Bering Sea has one of the largest marine mammal populations, possibly the largest clam population, one of the largest salmon runs, one of the highest densities of birds, and the largest eelgrass (*Zostera*) beds in the world; the yields of its pelagic and benthic commercial fisheries are extremely high.

Today the Bering Sea is the focus of more scientific research than at any time in its history. Three major efforts—the baseline studies sponsored by the Outer Continental Shelf Environmental Assessment Program of the Bureau of Land Management, U.S. Department of the Interior; intensified National Marine Fisheries efforts because of passage recently by the U.S. of the Fisheries Conservation and Management Act of 1977; and the International PROBES study sponsored by the Office of Polar Programs of the National Science Foundation—are now in progress. This treatise, therefore, might best be considered an interim progress report on the Bering Sea. In the immediate years ahead much more information will doubtless be published than is now available. It is hoped that this document will provide an historical baseline and point the way to further studies of this significant region of the world's oceans.

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The Eastern
Bering Sea Shelf:
Oceanography
and Resources

Section I

Physical Oceanography

Thomas H. Kinder, editor

A Perspective of Physical Oceanography in the Bering Sea, 1979

1

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ABSTRACT

Until recently, physical oceanographic research in the Bering Sea concentrated on broad spatial and long temporal scales, and much of the field work occurred off the shelf in water overlying the deep basins. Research concentrated on basin-wide phenomena of long duration, and this work determined the oceanic climate or physical geography of the Bering Sea.

Since about 1975, the focus of research has shifted toward shorter spatial and temporal scales, and also from the deep basins onto the shelf. Deviations from the large-scale mean state, such as interannual variability, fronts, eddies, tides, and vertical finestructure, are important biologically as well as physically, and this trend in research will probably continue through the next decade.

INTRODUCTION

About one-half of the Bering Sea overlies abyssal plain (depths $>3,500$ m), and about one-half overlies continental shelf (depths <200 m, Fig. 1-1). Until 1975 most physical oceanographic research concentrated on either the waters above the deep basins or those above the continental shelf near Bering Strait. This bias was illustrated in the symposium volume edited by Hood and Kelley (1974): physical oceanography papers concentrated on currents and water masses of the deep basins, and the monograph by Coachman, Aagaard, and Tripp (1975) focused on Bering Strait. With the exception of the Bering Strait region, little research was done on the continental shelf.

Much of the work before 1975 emphasized determination of mean conditions. Data were so few and so far apart, both in time and space, that variability could not be addressed rigorously. Investigators realized that smaller spatial and temporal scales were important, but they did not have adequate measure-

ments to define these scales. Hydrographic station spacing was often 100 km and stations were reoccupied only annually, if at all. Except for Bering Strait, direct current measurements were essentially nonexistent. Most work before 1975 aimed at establishing the mean state of the Bering Sea on large spatial scales, emphasizing the deep basin.

Since 1975 the focus has shifted from the longer temporal and broader spatial scales to shorter ones, and from the deep basins to the continental shelf. Investigations of interannual variations on broad spatial scales (climatic variability) and research in the deep basins continue, but the emphasis now lies elsewhere. Developments in instrumentation, trends in physical oceanography, and changes in funding have brought about this new emphasis.

The large spatial and temporal scales inherent in earlier work were primarily determined by the instrumentation and resources that were available. Hydrographic profilers (STD (salinity, temperature, depth) or CTD (conductivity, temperature, depth)) have increased vertical resolution from 10 m (or greater) to 1 m. Horizontal resolution was increased by generous amounts of shiptime dedicated to physical projects so that scales of 5 km or less were sampled. Satellite imagery permitted synoptic realizations of the surface thermal patterns and ice fields to scales of less than 1 km. Drogued drifting buoys tracked by satellite provided long tracks over large areas. Moored instruments measured currents to scales shorter than one hour. Sufficient shiptime was provided to repeat hydrographic surveys during a season, and to maintain current meter moorings for more than one year. Horizontal and vertical spatial resolution and temporal resolution were improved by one order of magnitude. Repeated measurements made it possible

to estimate persistence or change at these finer scales over the seasons and during different years.

Interest in these smaller scales was partly stimulated by growing theoretical and observational evidence in the oceanographic community that understanding broad-scale phenomena depended on understanding smaller-scale features. For instance eddies (MODE Group 1978), fronts (*J. Geophys. Res.* 83 (C9) 1978), finestructure (*J. Geophys. Res.* 83 (C6) 1978), and shelf dynamics (*Mémoires de la Société Royale des Sciences de Liège, X* 1976) were becoming increasingly important in the scientific literature. Renewed interest in shelf dynamics, overcoming a physical oceanographic bias toward deep water work, was partly the result of a new

perception. Before the mid-1979's physical oceanographers apparently believed that the continental shelves were rather chaotic, not amenable to study with standard oceanographic techniques, and in any case not as important as the deep ocean. Over the last decade we have found the shelves to be well organized (both hydrographically and dynamically), predisposed to fruitful study, and certainly important in their own right. In the Bering Sea, research on such topics as eddies, fronts, finestructure, and shelf dynamics naturally proceeded, based on the knowledge of broad-scale mean conditions established by earlier work.

The geographical shift was abetted by two large research programs: OCSEAP (Outer Continental

Figure 1-1. The Bering Sea: About one-half of the Bering Sea is abyssal plain at depths exceeding 3,500 m and about one-half continental shelf at depths less than 200 m. Recent work has concentrated on the shelf north of St. Lawrence Island and east of Nunivak Island/Pribilof Islands. This figure was prepared by Noel McGary for the atlas by Sayles et al. (1979).

Shelf Environmental Assessment Program, funded by the Bureau of Land Management), and PROBES (Processes and Resources of the Bering Sea Shelf, funded by the Polar Programs Division of the National Science Foundation). Both OCSEAP and PROBES have concentrated field work on the shelf, and together the resources available for shelf research have increased tenfold or more. OCSEAP aims for a broad understanding of the shelf to assess hazards of petroleum development, while PROBES aims to understand the southeastern shelf ecosystem by concentrating on the trophic web of the Alaska or walleye pollock (*Theragra chalcogramma*). Justification for such large research programs was the realization of man's impact on coastal environments and of the importance of the continental shelves and their waters to man. In the Bering Sea, the potential petroleum resource and the renewable fisheries resource stimulated research requirements and subsequent funding.

The seven chapters in this section reflect both the foundation provided by the earlier work and the direction of research that the influence of new instrumentation, new ideas, and new funding has produced. Work has continued in the deep Bering Sea, and I will review the progress there before outlining the physical oceanography section of this book. Then, in the light of the work recently completed, I will discuss some research needs and probable progress in Bering Sea oceanography in the next decade.

DEEP BERING SEA

By the time of the Hood and Kelley (1974) volume the patterns of the mean circulation and the mean hydrographic distributions were established, and something about their annual and interannual variations was known. Hughes et al. (1974), Favorite (1974), Favorite and Ingraham (1974), Favorite et al. (1976), Ohtani (1973), and Arsenev (1967) showed a basically cyclonic circulation which strengthens considerably in winter (Fig. 1-2). Similarly, the basic hydrographic state was detailed by Arsenev (1967), Ohtani (1973), and Favorite et al. (1976). Establishing these mean states was important. From the salinity distribution in Fig. 1-3 we can correctly infer a cyclonic circulation over the deep basin, a haline front over the northeastern continental slope, and a strong riverine influence near shore. Nevertheless figures like 1-2 and 1-3 present a picture that is deceptively smooth spatially (and by implication temporally). Recent work, based on improved resolution, has concentrated

Figure 1-2. A circulation scheme proposed by Hughes et al. (1974); the numbers refer to measured speeds (double arrows) in cm/sec. The general cyclonic circulation over the deep basins, the strong southward flow adjacent to the western boundary (East Kamchatka Current), and the northwestward flow paralleling the shelf break (Bering Slope Current), are well founded. Observations and modeling indicate that bottom features such as Shirshov and Bowers ridges influence flow, and that the small gyres suggested here are probably transient eddies. Considering the evidence then available, they prudently left the shelf blank (arrows could have been safely added in the Bering Strait).

on the spatial and temporal variability (i.e., deviations from mean conditions).

Sayles et al. (1979) used volumetric analysis to examine the annual cycle of water mass properties over the deep basin. Assessing seasonal changes in temperature and salinity against the background of mean conditions, they produced an atlas of water mass properties in the upper 1,500 m of the basin. This atlas, focused on the deep basin and on annual periods, is the culmination of earlier work directed at large spatial scales and long temporal scales.

There have been two efforts to model the Bering Sea circulation using wind stress and inflow conditions as forcing. Bacon (1973) used a barotropic model of the deep basin with 56-km grid spacing, and Han and Galt (1978) used a similar model covering both deep basin and shelf with 100-km grid spacing. The results of both models over the deep basin were reasonable—a general cyclonic circulation modified by large bathymetric features, and modulated by the annual cycle of wind stress. Over the shelf the Han and Galt model showed flow toward Bering Strait concentrated on the western side of the shelf, in accord with present views. These models are not fully realistic in other respects (as their authors point

Figure 1-3. Mean surface salinity for summer (g/kg, taken from Favorite et al. 1976). This pattern not only conforms to the general circulation over the deep basins, but suggests the front overlying the slope. The mean horizontal salinity gradient, which is essentially congruent with the horizontal density gradient, is an important driving force for flow over the slope and over parts of the shelf. The focus of research is now shifting from such long-term and broad-scale distributions as shown here (and in Fig. 1-2), toward departures from the mean at many temporal and spatial scales. Research topics include climatic variability, tidal patterns, mesoscale eddies, fronts, and finestructure. Note that the intervals between isohalines are not uniform.

out). Over the shelf important features such as tidal flow, fronts, and stratification are not included in the larger area model. Over the deep basin these models are incapable of resolving the ubiquitous mesoscale (about 50-100 km diameter) eddies found there. Modeling promises to increase our understanding in the future, but this promise is not yet fulfilled.

Recent observational work in the deep basin has concentrated on eddies (i.e., spatial variability) with scales of 10-100 km. Kinder et al. (1975) surveyed the Bering Slope Current (Arsenev's 1967 Transverse Current). While it flowed northwestward parallel to the shelf break, a series of eddies about 100 km in diameter were embedded in the current. Kinder and Coachman (1977) found a similar-sized eddy in outer Pribilof Canyon, but inferred that its generating mechanism differed from those seen earlier. Kinder et al. (1980) detected an eddy in the southeastern corner of the Aleutian Basin by using drogued drifters tracked by satellite and measured the eddy by hydrographic survey. A second survey of the area eight months after the eddy was detected showed that it had dissipated or moved. These three papers focused

on the transitory nature of the eddies (rather than the alleged permanence of similar sized gyres or "gyrals" in some previous works), but the most startling evidence of variability came from satellite observation. Solomon and Ahlnäs (1978) showed patterns of surface temperature and ice flows that revealed a closely packed field of eddies adjacent to the Kamchatka Peninsula. Unlike the earlier observations of eddies enumerated in the previous paragraph, the eddies that Solomon and Ahlnäs detected with satellite infrared and visual imagery did not resemble typical open-ocean eddies or rings. Instead, the Kamchatka eddies had spiral arms and were closely packed in an eddy field.

One further smaller-scale study appeared after 1974. Swift and Agaard (1976) used closely spaced (~10 km) hydrographic stations near Samalga Pass (about 200 km west of Unimak Pass) to study the upwelling or mixing that occurs along the Aleutian Islands. They concluded that the data supported upwelling, a process (along with turbulent mixing) that may be important along the length of the Aleutian Chain.

Thus work in the deep basin continues and, like work on the shelf, it focuses more on variability than in the past. Progress has also continued in another direction, that of assessing more finely the oceanic and atmospheric climate of the Bering Sea, and the effects of departures from this climate (see Ingraham, this volume). Further progress using both climatological and smaller-scale points of view should improve understanding of the shelf oceanography, because the waters above the basin interact with those above the shelf (see Coachman and Charnell 1979).

BERING SEA CONTINENTAL SHELF

The seven chapters in this section provide a sampling of recent Bering Sea work. We attempted to synthesize earlier work, and to present a broad perspective of regional processes. At least four of the chapters are preliminary reports, because the authors are currently working on extending the results reported here. All chapters have been reviewed by outside researchers (usually anonymous), and we very much appreciate their advice and assistance.

Chapter 2, Climate

The Bering Sea undergoes many changes annually in such measures of climate as air temperature, mean wind speed, incidence of storms, river runoff, and extent of ice cover. These changes affect the hydro-

graphy and the circulation of the shelf waters. J. E. Overland describes some of these important climatic elements.

Chapter 3, Climatic variability

Weather patterns change from year to year, deviating to a greater or lesser extent from long-term means. H. J. Niebauer presents a case study of the effects of 700mb atmospheric flow deviations on sea surface temperature.

Chapter 4, Southeastern hydrography

The southeastern shelf can be separated into three distinct hydrographic domains based on vertical structure (i.e., stratification). The boundaries between these domains are fronts, each front with different properties. T. H. Kinder and J. D. Schumacher discuss this hydrographic structure as well as features such as finestructure and upwelling.

Chapter 5, Southeastern circulation

Most of the flow over the southeastern shelf is tidal, but there is significant subtidal (low-frequency) and mean flow. The characteristics of the mean and low-frequency flow can be separated into three flow regimes that are nearly coincident geographically with the hydrographic domains. T. H. Kinder and J. D. Schumacher describe the character of this circulation and the resulting flow regimes.

Chapter 6, Norton Sound

The most isolated large embayment in the eastern Bering Sea, this shallow sound has hydrography and circulation similar to the entire shelf, although there are important differences. It is influenced by tides, winds, and river runoff (Yukon River), but also by the exterior flow toward Bering Strait. R. D. Muench, R. B. Tripp, and J. D. Cline describe the physical oceanography of this interesting corner of the shelf.

Chapter 7, Bering Strait

This connection between the Bering and Chukchi Seas affects both Bering Sea shelf dynamics and the oceanographic state of the Arctic Ocean. Mean flow through this shallow and narrow strait is northward, but the flow is distinctly unsteady. L. K. Coachman and K. Aagaard discuss the flow through Bering Strait with emphasis on low-frequency variability observed in current-meter data, and suggest a lower value for mean transport than previously proposed.

Chapter 8, Tides

Most of the kinetic energy over the shelf is tidal, and this energy profoundly affects stratification and stirring. Furthermore, the Bering Sea has been identified as important in the global dissipation of tidal energy. C. A. Pearson, H. O. Mofjeld, and R. B. Tripp discuss recent tidal observations and present charts of tidal heights and tidal currents for the eastern shelf.

Even taken together these chapters are not yet a coherent and complete synthesis of the physical oceanography of the Bering Sea shelf. They are a major step toward the goal, however, and a measure of progress is that no physical studies in the Hood and Kelley volume focused on the shelf. The references in the chapters of this section demonstrate that these chapters in fact represent over fifteen shelf papers published since 1974, and more are in preparation.

FUTURE STUDY

Any prediction of future trends or probable results would soon be overtaken by events, but I can list areas of present research and the needs within these areas for the next ten years or so. The list is my own, but I judge that it is nearly a consensus. Over the next decade research should further analyze data already collected, extend geographical coverage, relate local findings to general physical processes, and incorporate physical results into ecosystem understanding.

Data analysis

Immense quantities of data have been collected and, as this volume illustrates, they have yielded interesting results. Much can still be learned from the data, however. In particular, the chapters by Kinder and Schumacher (Circulation, 5), Coachman and Aagaard (Bering Strait, 7), and Pearson, Mofjeld, and Tripp (Tides, 8) are distinctly preliminary reports. The ideas in these chapters can be more fully developed with existing data, and other investigators also have ideas that can use already available data. OCSEAP has emphasized data collection more than data analysis and scientific synthesis, so that there is now a backlog of data undigested by the researchers.

Geographical coverage

Collection of physical data has been intensive over the southeastern shelf (Kinder and Schumacher, Chapters 4 and 5) and near Bering Strait (Muench

et al., Chapter 6; Coachman and Aagaard, Chapter 7). Between St. Lawrence Island and Nunivak Island on the eastern shelf there has been sparse coverage, and northwest of the Pribilof Islands no coverage.¹ To some extent the entire shelf behaves as a single system, and our local understanding (e.g., at a particular lease area) depends on our understanding of the entire shelf system. The region between St. Lawrence and Nunivak islands connects the two best-understood regions, and the western shelf probably provides most of the water to the northward outflow in Bering Strait. A complete tidal picture (Pearson et al., Chapter 8) cannot be painted without coverage across the entire shelf. We will not be able to construct a comprehensive description of physical processes on the shelf nor understand their workings until geographic coverage is complete. At the same time, extending geographical coverage should be done based on results of previous work, both recent and early. Extended coverage should not take the form of broad reconnaissance surveys, but of experiments with well-defined objectives.

Physical processes

Many phenomena revealed in Bering Sea data illustrate and identify processes that have wider than regional importance. Reports of these findings should be made in the oceanographic literature, and this has begun. Some of these processes and phenomena include:

Finestructure. Inversions in vertical temperature and salinity profiles on scales of 1-10 m have been found over the outer southeastern shelf, showing seaward intrusions of shelf water and shoreward intrusions of oceanic water (Coachman and Charnell 1977, 1979). Such finestructure affects mixing on the shelf, and observations here add to an understanding of oceanic mixing.

Shelf dynamics. Interest in flow on continental shelves has grown rapidly over the past five years, and the Bering Sea shelf has unusual properties (e.g., breadth, ice cover, shape of shoreline) that make our observations particularly useful. Initial reports on shelf circulation and dynamics appear in this volume (Kinder and Schumacher, Chapter 5; Muench et al. Chapter 6; Coachman and Aagaard, Chapter 7), and additional reports should be forthcoming in the next few years.

Tides. Most of the horizontal kinetic energy over the Bering shelf is tidal, and it is thought that dissipation of the global tidal energy occurs in shallow seas. Understanding tides on the Bering shelf is important locally, but also for understanding tidal behavior in shallow water generally. Observations (Pearson et al., Chapter 8) and numerical modeling (Leendertse and Liu 1980) are increasing this understanding.²

Air-sea interaction. Strong annual signals in atmospheric forcing occur over the Bering shelf, and the shelf waters respond in hydrography, circulation, and ice cover. Because of the strength of annual change, this shelf is a natural atmosphere-ocean interaction laboratory (Reed 1978, Muench and Ahls 1976, Muench and Charnell 1977).

Fronts. Sharp transitions of many sizes have been discovered throughout the oceans and in shallow seas. Across the southeastern shelf three fronts have been identified, separating regions with distinct hydrographic, circulation, biological, and sedimentary properties. These fronts clearly have dynamical and mixing implications, and understanding fronts is important for understanding the Bering shelf and the oceans in general. (Kinder and Coachman 1978, Coachman and Charnell 1979, Schumacher et al. 1979, Iverson et al. 1980).

Eddies. Variability on scales of a few tens to a few hundreds of kilometers has been increasingly recognized as important over the past decade. While no eddies have yet been discovered on the Bering Sea shelf, they frequently inhabit waters adjacent to the shelf and they may influence cross-shelf exchange and mixing (Kinder et al. 1975; Kinder and Coachman 1977; Kinder et al. 1980).

Ecosystem studies

In addition to physical studies, much work in chemical, biological, fisheries, and geological oceanography has been done concurrently. Within data collected by many disparate groups are clues to relationships between biota and environment. PROBES investigations have begun to examine the shelf as an ecosystem (Iverson et al. 1979, Iverson et al. 1980) by focusing on the walleye pollock food web. Better understanding of the interrelationships between components of the ecosystem occurs not only through studying apparent links between components, but

¹ There are many historical hydrographic stations, but their usefulness is limited. These data are typically widely separated vertically, horizontally, and temporally.

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through studying the components themselves. For example, if physical oceanographers better understood fronts and how they affect mixing, then the ecology of Bering shelf plankton would be more intelligible. It does not seem too optimistic to believe that PROBES and OCSEAP work can have a synergistic relationship, both benefiting from different points of view.

CONCLUDING REMARKS

I have enumerated various areas of study that I believe fruitful, and there is strong motivation to do these studies. I separate these motivations into non-scientific, physical oceanographic, and multidisciplinary categories.

I assume that with more scientific understanding of the shelf, man can better exploit and manage its resources. Two obvious resources are the potential petroleum reservoirs and the large fish stocks (see Fisheries Oceanography and Fisheries Biology Sections). A further resource is less tangible, but may be best illustrated by considering marine mammals. Many citizens have a strong moral feeling that protecting and preserving wildlife and the natural environment is important apart from economic profit, and this feeling has been put into law. Extracting oil economically, maintaining maximum sustainable fisheries, and preserving wildlife all require knowledge of the Bering shelf environment and the desire for these ends provides a general motivation for further study.

In physical oceanography, the knowledge already gained of unique regional features motivates continued work. Present physical understanding, as illustrated in this section of the book, provides a basis for future scientific success. For example, we now have a general description of the inner front, so that we can design an experiment to study frontal dynamics and cross-frontal mixing. This shelf also has unique characteristics, so that studies in the Bering Sea may isolate important features that are obscured on other shelves. The unusual width, exceeding 500 km, has made the hydrographic domains and frontal systems clear; similar tendencies on narrower shelves may be too compressed to show clearly. The absence of a strong boundary current at the shelf break permits shelf processes to dominate; over the shelf of the southeastern United States, for example, the Gulf Stream strongly influences shelf processes. Strong annual variations, as in the extent of ice, permit examining effects by comparison (e.g., ice-covered versus ice-free); by contrast, the broad shelf in the Barents Sea may have ice throughout the year. The

outflow at Bering Strait forces a strong cross-isobath flow across the shelf; most shelves have only a weak cross-isobath flow caused by river runoff.³ Thus there are sound reasons for continuing physical studies on the Bering shelf.

Some of the most exciting recent results have been biological, and these have resulted from a multidisciplinary outlook. Capitalizing on previous fisheries research and the physical understanding produced by OCSEAP and earlier work, PROBES investigators have found that the shelf ecosystem is organized by physical phenomena. For example, the distribution of phytoplankton species resembles the distribution of hydrographic domains. Because we now have a basic description of physical processes over much of the shelf, non-physical disciplines can better understand their data and non-physical data can confirm, modify, or refute physical ideas. Multidisciplinary process studies can now use the foundation of present knowledge to elucidate important aspects of the shelf ecosystem. For instance, the frontal dynamics experiment suggested in the preceding paragraph would benefit from non-physical measurements: how are plankton, murre, and methane affected by the front? Even the simple expedient of concurrent sampling (e.g., hydrographic casts and chlorophyll sampling) can increase the value of independent sampling and use shiptime more efficiently.

In conclusion, the geographic phase of Bering Sea oceanography has ended and mean conditions on broad scales are known. Present interest is in the details of spatial and temporal variability on many scales—from meters to hundreds of kilometers, and from hours to decades. In physical oceanography fronts, eddies, and waves have largely supplanted water masses, gyres, and currents as pervasive objects of research. Investigators have begun to digest massive amounts of data collected during the last five years, and important results of their analyses will be forthcoming not only with a regional perspective, but with both narrow disciplinary and broad multidisciplinary views. Sound motivation exists for continued study of the Bering Sea continental shelf.

ACKNOWLEDGMENTS

L. K. Coachman, F. Favorite, M. A. Sayles, J. H. Swift, and R. B. Tripp read an early draft of this paper and made many helpful comments. L. K. Coachman, J. D. Schumacher, and R. B. Tripp have been my close scientific colleagues in Bering Sea work, and any valid insights that I have are at least

³ At peak discharges the Yukon River contributes only 2 percent of the Bering Strait outflow.

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CONCLUDING REMARKS

I have enumerated various areas of study that I believe fruitful, and there is strong motivation to do these studies. I separate these motivations into non-scientific, physical oceanographic, and multidisciplinary categories.

I assume that with more scientific understanding of the shelf, man can better exploit and manage its resources. Two obvious resources are the potential petroleum reservoirs and the large fish stocks (see Fisheries Oceanography and Fisheries Biology Sections). A further resource is less tangible, but may be best illustrated by considering marine mammals. Many citizens have a strong moral feeling that protecting and preserving wildlife and the natural environment is important apart from economic profit, and this feeling has been put into law. Extracting oil economically, maintaining maximum sustainable fisheries, and preserving wildlife all require knowledge of the Bering shelf environment and the desire for these ends provides a general motivation for further study.

In physical oceanography, the knowledge already gained of unique regional features motivates continued work. Present physical understanding, as illustrated in this section of the book, provides a basis for future scientific success. For example, we now have a general description of the inner front, so that we can design an experiment to study frontal dynamics and cross-frontal mixing. This shelf also has unique characteristics, so that studies in the Bering Sea may isolate important features that are obscured on other shelves. The unusual width, exceeding 500 km, has made the hydrographic domains and frontal systems clear; similar tendencies on narrower shelves may be too compressed to show clearly. The absence of a strong boundary current at the shelf break permits shelf processes to dominate; over the shelf of the southeastern United States, for example, the Gulf Stream strongly influences shelf processes. Strong annual variations, as in the extent of ice, permit examining effects by comparison (e.g., ice-covered versus ice-free); by contrast, the broad shelf in the Barents Sea may have ice throughout the year. The

outflow at Bering Strait forces a strong cross-isobath flow across the shelf; most shelves have only a weak cross-isobath flow caused by river runoff.³ Thus there are sound reasons for continuing physical studies on the Bering shelf.

Some of the most exciting recent results have been biological, and these have resulted from a multidisciplinary outlook. Capitalizing on previous fisheries research and the physical understanding produced by OCSEAP and earlier work, PROBES investigators have found that the shelf ecosystem is organized by physical phenomena. For example, the distribution of phytoplankton species resembles the distribution of hydrographic domains. Because we now have a basic description of physical processes over much of the shelf, non-physical disciplines can better understand their data and non-physical data can confirm, modify, or refute physical ideas. Multidisciplinary process studies can now use the foundation of present knowledge to elucidate important aspects of the shelf ecosystem. For instance, the frontal dynamics experiment suggested in the preceding paragraph would benefit from non-physical measurements: how are plankton, murre, and methane affected by the front? Even the simple expedient of concurrent sampling (e.g., hydrographic casts and chlorophyll sampling) can increase the value of independent sampling and use shiptime more efficiently.

In conclusion, the geographic phase of Bering Sea oceanography has ended and mean conditions on broad scales are known. Present interest is in the details of spatial and temporal variability on many scales—from meters to hundreds of kilometers, and from hours to decades. In physical oceanography fronts, eddies, and waves have largely supplanted water masses, gyres, and currents as pervasive objects of research. Investigators have begun to digest massive amounts of data collected during the last five years, and important results of their analyses will be forthcoming not only with a regional perspective, but with both narrow disciplinary and broad multidisciplinary views. Sound motivation exists for continued study of the Bering Sea continental shelf.

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³ At peak discharges the Yukon River contributes only 2 percent of the Bering Strait outflow.

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Marine Climatology of the Bering Sea

2

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ABSTRACT

The climate of the Bering Sea is strongly related to the presence and movement of marginal sea ice. In winter, weather elements are continental and arctic in character, being replaced by maritime influences from the south in summer. In winter this results in north to easterly winds,¹ a tendency for clear skies, and substantial diurnal temperature range. Summer is characterized by a progression of storms through the Bering rather than fixed weather types, producing increased cloudiness, reduced diurnal temperature range, and winds rotating through the compass with a slight tendency for southwest.

INTRODUCTION

The climate of the major portion of the Bering Sea can be classified as "polar oceanic." The system of Köppen and Geiger (1930) classifies "polar" as a geographic region in which the mean monthly temperature for the warmest month is 10 C or less. This is valid for all but the inner region of Bristol Bay. Estienne and Godard (1970) further divided Köppen's classification into five subclasses. The "polar oceanic" region is described as being humid with annual precipitation of from 300 to 500 mm, fairly uniformly distributed over the seasons. This contrasts with the Beaufort Sea which is "polar with a large range of temperature." The Beaufort Sea has less annual precipitation than polar oceanic and greater temperature and precipitation contrasts between summer and winter.

The Bering Sea is affected by arctic, continental, and maritime air masses. In summer the entire region is normally under the influence of maritime air from the Pacific. The southern portion of the Bering Sea is most frequently under the influence of maritime air, except for January and February (Grubbs and McColum 1968), when normally a strong flow of air from

the north and east brings continental and arctic air to most of the area. Arctic air also persists in spring and fall in the northern sector. Using the predominance of the arctic high-pressure air mass over the northern Bering Sea as the indicator, the winter circulation pattern persists for nine months, September through May.

A major influence of the general circulation on the area is the region of low pressure normally located in the vicinity of the Aleutian chain, referred to as the Aleutian Low. On monthly mean pressure charts this appears as a low-pressure cell normally oriented with the major axis in an east-west direction. This is a statistical low, indicating only that pressures are generally lower along the major axis as a result of the passage of low-pressure centers or storms. Storms are most frequent in this area and are more intense there than in adjacent regions. The most frequent track or trajectory of movement of these storms is along the Aleutian Islands and into the Gulf of Alaska in winter, and along the same general path in the west but curving northward into the Bering Sea in summer. The monthly frequency of low-pressure centers in the southern Bering Sea is slightly higher in winter (generally four to five) than in summer (three to four). However, winter storms are much more intense.

In winter the most frequent airflow is north-easterly around the northern side of the low-pressure cell present at some location along the Aleutian chain. In summer, with the movement of lows into the Bering Sea, a more southwesterly mean flow develops over the lower two-thirds of the region. In the Aleutian chain in winter and during the summer in the southern Bering Sea, frontal activity can be severe as very cold arctic or continental air comes in contact with the warm air from the Pacific Ocean, forming a sharp discontinuity. In the northern Bering

¹ Wind convention names the direction from which the wind is blowing.

Figure 2-1. Location map for the eastern Bering Sea.

Figure 2-2. Annual temperature march and monthly standard deviations for Northeast Cape and St. Paul. There is greater annual range at the northern station.

Sea, polar air masses usually predominate for extended periods of time. Frontal systems moving through the area generally represent a line of discontinuity of air masses of similar character, and consequently are much less severe.

The ice pack dominates the Bering Sea from January through May (McNutt and Pease, this volume). After April the ice pack moves progressively northward and by July it is generally north of Bering Strait. The Bering Sea is ice free from July through September. Ice begins moving southward during October and reaches its southern limit near 60°N by February or March. The seasonal migration of the ice pack is extremely important for the climate of the Bering Sea. It introduces a continental influence during winter which allows cold arctic and continental air to establish itself over the ice-covered sea with wide ranges in daily and seasonal temperatures. This contrasts sharply with the entire Bering Sea in summer and the southern portion in winter, which have a maritime climate with a more uniform daily temperature regime and enhanced precipitation.

CLIMATOLOGICAL SUMMARY

Data for this section were obtained from *Climatic Atlas of the Outer Continental Shelf Waters and Coastal Regions of Alaska: Volume II—Bering Sea* (Brower et al. 1977) and *A Climatological Guide to Alaskan Weather* (Grubbs and McCollum 1968). The data periods in Grubbs and McCollum "contain at least 10 years" while Brower et al. specify the number of observations. Fig. 2-1 is a map locating stations and marine areas used in the figures and tables. Marine areas A to E designate regions where ship reports form the climatological base.

Temperature

Since this region is oriented north-south, there is some latitudinal variation in temperature. This is more noticeable in winter when the total amount of sunshine varies from several hours in the south to almost none in the north. In addition the southern portion in winter may come under the influence of southeasterly flow of fairly warm, moist air from the north Pacific, while the northern portion is under the influence of cold, dry air generally flowing out of the interior of northern Alaska. While the ice serves as a barrier between air and sea, some heat is diffused through the ice from the much warmer ocean so that, although the arctic air is very cold, temperature minimums do not reach the low extremes of the Alaskan interior.

Fig. 2-2 plots the mean monthly temperatures and

standard deviations for Northeast Cape (approximately 3,000 observations per monthly group) and for St. Paul (over 4,000 observations per monthly group) after Brower et al. (1977). Fig. 2-2 shows much larger annual range in the north. The August-minus-February temperature difference is 25 C, versus a 14 C difference at St. Paul. The standard deviation of monthly temperatures at Northeast Cape is greater in winter than in summer and greater than the standard deviations for all seasons at southern stations. Table 2-1 presents the mean maximum and

minimum temperatures ($^{\circ}\text{C}$) by month for Northeast Cape and Adak after Grubbs and McCollum (1968). When the mean daily ranges of temperatures (monthly mean maximum minus monthly mean minimum) for summer and winter are compared, Northeast Cape shows an 8 C range in late winter and a 4 C range in summer, while the range at Adak is 4-5 C, independent of season. The greater temperature ranges at Northeast Cape point to a continentality in the north during winter which is replaced by a maritime climate in summer.

TABLE 2-1

Maximum and minimum mean monthly temperatures and monthly precipitation
for Northeast Cape and Adak

	Temperatures in °C					Precipitation in mm						
	NORTHEAST CAPE											
	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
Mean maximum temperature	−12	−13	−9	−6	2	7	11	10	7	2	−3	−13
Mean temperature	−16	−17	−13	−9	−31	4	8	8	5	0	−4	−15
Mean minimum temperature	−19	−21	−17	−13	−3	1	6	6	3	−2	−7	−17
Mean total precipitation	12.9	10.2	21.1	9.4	13.7	12.9	28.2	98.8	113.0	56.4	39.9	14.5
	ADAK											
	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
Mean maximum temperature	3	3	4	6	7	9	12	13	11	8	5	3
Mean temperature	1	1	2	3	5	7	9	11	9	6	3	1
Mean minimum temperature	−1	−2	−1	1	3	5	7	8	7	3	1	−1
Mean total precipitation	160.0	134.6	165.1	106.7	116.8	83.8	76.2	104.1	137.2	185.4	193.0	198.1

Precipitation and cloud cover

There is a general decrease in the amount of precipitation from south to north because the northern points are more distant from the moisture source, especially in winter. Table 2-1 shows that precipitation at Northeast Cape is low from December through June, when the northern region is dominated by the arctic air mass. There is a sharp spike during August and September, when storm tracks penetrate the northern Bering Sea. Adak has precipitation the year round with an increase in October through December which corresponds to a season of cyclogenesis in the southern Bering Sea.

One of the most important controls effected by cloudiness is of the type of air which is synoptically in possession of the area. In winter the majority of the Bering Sea region is most frequently under northeasterly or northerly flow of cold, dry arctic air. In summer the entire region is under the influence of moist air from a north Pacific air mass. This leads to a larger number of clear days in the northern region in winter. However, even though the arctic air contains only a small amount of moisture, the cold air mass exhibits high relative humidities near the surface. Only a slight amount of lift, for example from a weak storm system, is required for formation of cloud cover. Fig. 2-3 shows the percent of observations by month in which the observed cloud cover was five-eighths or more for Northeast Cape, St. Paul, and Adak (Brower et al. 1977).

Figure 2-3. Percent of observations reporting five-eighths cloud cover or greater.

Figure 2-4. February wind roses. (Direction from which the wind is blowing.)

Surface Winds

Fig. 2-4 shows wind roses for selected locations in the Bering Sea during February and Fig. 2-5 shows roses for the same stations during August. Wind roses show the percentage of observations from each of eight possible directions. The number in the center is the percent of light winds in the record. Each arm is divided into the percent of observations of 1-6 kn, 7-16 kn, and 17 kn or greater from each direction.

In winter the northern stations show a high percentage of winds greater than 17 kn from the north and northeast, while in the south, the winds over marine area C are uniformly distributed over direction with moderately large speeds. This marine rose is indicative of a fairly continuous progression of storms through the area. Wind speeds over the Bering Sea in summer are generally lower than in winter,

Figure 2-5. August wind roses. (Direction from which the wind is blowing.)

although conditions are seldom calm. Marine area A to the north shows little preferred direction, but the other stations show predominance of south and southwest winds.

Runoff

Freshwater inflow to the eastern Bering Sea is primarily from the Yukon, with only minor contributions from other sources, most notably the Kuskokwim. Fig. 2-6 plots the mean monthly discharge rates for the combined Alaska runoff into the Bering Sea and for the Yukon for 10 water-years 1967-77 (U.S.G.S. 1977). The 10-year monthly mean discharge rates show very little flow and variability of flow during the months of December, January, February, March, and April. The months of greatest flow are also the months of greatest variability: May and June.

CIRCULATION

There are two general approaches to classifying climatological types: a synoptic climatology which regards circulation patterns as an implicit function of the static sea level pressure (SLP) distribution (Barry and Perry 1973), and a kinematic approach in which synoptic weather maps are classified in terms of principal storm tracks (Klein 1957).

Two synoptic climatologies which refer to the Bering Sea are those of Putnins (1966) and Barry (1978). Putnins establishes 22 patterns "in such a way that for every date of this period (1 January 1945 to 31 March 1963) a specific baric weather pattern could be assigned." Unfortunately, Putnins' emphasis centers on continental Alaska and applies only in a general manner to the Bering Sea. Barry developed a synoptic climatology for the Chukchi Sea which has 22 types and includes the maritime region of northern Bering Sea.

Barry states that in winter his Type 1, arctic high pressure with subpolar easterlies at Kotzebue (Fig. 2-7) is dominant and is associated with a low-level atmospheric temperature inversion. Interruptions by anticyclonic systems are the most common. They are associated with cold continental air masses which reinforce the shallow arctic boundary layer. Cyclonic interruptions are less common

Figure 2-6. Runoff rate as a function of month for the Yukon River and for total runoff into the eastern Bering Sea.

Figure 2-7. Barry's (1978) dominant winter Type 1 sea-level pressure weather pattern.

and bring warm air with larger amounts of cloudiness. Barry states that in summer there is a great variety of types, with both cyclonic and anticyclonic types apparent. Type 3 (Fig. 2-8) is the most common in July and is closest to the mean monthly SLP charts for summer given by Brower et al. (1977); one interpretation is that atmospheric circulation in the

Figure 2-8. A frequently occurring summer weather type, from Barry (1978).

southern Bering Sea can be more readily characterized by mean storm tracks or the presence of low-pressure centers in certain sectors of the region than by static weather types. Fig. 2-9 plots the average number of low-pressure centers observed in a $10^\circ \times 10^\circ$ latitude-longitude area during the nine-year period of record 1966-74. These areas are NW (60° - 70° N, 170° - 180° W), NE (60° - 70° N, 160° - 170° W), SW (50° - 60° N, 170° - 180° W), and SE (50° - 60° N, 160° - 170° W). It is apparent that southern sectors have two to three times more storms than the northern sectors. However, the monthly variability is high, suggesting that the period of record is too short to make comparisons between months.

Figure 2-9. Frequency of low-pressure systems by month for the northern and southern Bering Sea.

CLOUD STREETS

The advection of cold air southward from the north and northeast Bering Sea in winter produces ideal conditions for convective cloud development over the relatively warm waters to the south. The air is virtually unimpeded as it flows south across the ice, and the ice edge forms a sharp line of demarcation where sea temperatures can be as much as 15 C warmer than the air. As the air continues to flow south it is progressively destabilized by the upward transfer of heat and moisture from the ocean.

The most frequently observed patterns are of the type displayed in Fig. 2-10, which shows uniform cloud streets at intervals averaging 5-6 km forming 20-70 km to the south of the ice edge and aligned in the direction of the surface wind (Streten 1975). They extend some 200-300 km downstream, displaying only a small increase in cloud element dimensions.

Figure 2-10. Satellite photo from 23 March 1978 showing cloud streets forming south of the ice sheet. Note Alaska to the right.

Beyond this distance there is a sharp transition from parallel streets to an open cell convection pattern with cloud elements 10 km in width separated by 20-25 km. With increasing distance from the source, the open areas grow to 50 km with cells of 20 km.

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Recent Short-period Wintertime Climatic Fluctuations and Their Effect on Sea-surface Temperatures in the Eastern Bering Sea

3

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ABSTRACT

Upper air (700 mb) winter pressure patterns have shown sharp fluctuations over the period 1963-78. Mean annual sea-surface temperature (SST) fluctuations appear to be an effect of these short-term climatic fluctuations. The mid-1960's were a time of southerly flow of air leading to above-normal SST. A rather sharp reversal in atmospheric conditions led to a sharp drop in SST in the early to middle 1970's. Since 1977, the upper air flow has become southerly, leading to a sharp rise in SST. Autocorrelation analysis of the SST suggests that these trends persist for at least two years.

INTRODUCTION

Since the early 1960's there have been several unusually strong short-term climatic fluctuations (time scale of one to several years) over the Northern Hemisphere. The most recent of these fluctuations is best illustrated by the severe North American winters of 1976-79. However, this latest fluctuation has also caused abnormally warm winters in Alaska and the adjacent areas. This chapter examines these short-term climatic variations and their relationship to fluctuations in sea-surface temperature (SST) in the eastern Bering Sea.

Favorite and McLain (1973) did a computer analysis of two million observations of SST for the North Pacific Ocean for 1953-60. Evidence is given for an orderly transpacific movement of extensive warm and cold surface-water anomalies. They suggest that these features move about the North Pacific gyre with periods of five to six years and profoundly affect environmental conditions off the West Coast of the United States and Canada. These fluctuating

environmental conditions are probably related to fluctuations in SST in the Bering Sea discussed in this chapter.

The eastern Bering Sea SSTs in the mid-1960's were generally above normal (Niebauer 1978). Johnson and Seckel (1976) pointed to a climatic shift which occurred in the early 1970's with drastic effect on some of the Bering Sea fisheries. They cited low salmon catches in 1973 and 1974 as attributable to the unusually cold winters of 1971 and 1972. McLain and Favorite (1976) related the cold SST to large-scale changes in the atmospheric circulation which caused increased northerly winds over the Bering Sea. The onset of the decline in SST coincided with anomalous southward extent of the ice pack (Kukla and Kukla 1974). Fluctuations in the southern ice extent in the Bering Sea in relation to the short-term climatic fluctuations are considered in Chapter 9 of this volume.

More recently, Niebauer (1980) has related a subsequent sharp rise in SST in this region in 1975-78 again to changes in the atmospheric circulation which caused southerly flow over the Bering Sea. Namias (1978) has suggested that the latitudinal SST patterns across the North Pacific in November 1976 foretold the strong and persistent air flow from the south over the Bering Sea during the winter of 1976-77. This may be related to the transpacific movement of warm and cold pools of water noted by Favorite and McLain (1973). This trend has persisted and intensified through the winter of 1978-79 (Niebauer 1980). The following sections of this chapter consider a description of the upper-air (700 mb or 3,000 m)

flow for the winters of 1962-79. The patterns are related to a 16-year time series of SST for the eastern Bering Sea.

DATA

Northern Hemisphere monthly 700 mb pressure charts from the period 1963-79 from *Monthly Weather Review* were analyzed. Representative examples were chosen to illustrate the salient points in the following sections. Sea-surface temperatures (SST) were obtained from the Naval Fleet Numerical Weather Central, Monterey, California, through D. R. McLain of the National Marine Fisheries Service. The SSTs are 12-hourly analyses of observations from ships of opportunity in the eastern Bering Sea. These observations have been arithmetically averaged to give the mean annual time series in Fig. 3-1.

There are, at times, strong horizontal temperature gradients near the Pribilof Islands and in Bristol Bay that may bias the SST data. In addition, in years of above-normal ice extent, ships are forced out of portions of Bristol Bay and to the south of the

Pribilof Islands. This may bias the SST to warmer temperatures. An example of this is a comparison of the SSTs for 1975 and 1976. In general, 1976 was a colder year (Niebauer 1980) and yet the SST was ~ 0.3 - 0.5 C higher (Fig. 3-1). However, shelf-wide bottom temperatures for June 1975 and 1976 (Fig. 9-3, Niebauer, Chapter 9, this volume) were nearly identical. The greater ice extent in 1976 relative to 1975 may have been responsible for the slightly higher temperature bias in 1976. Although there may be occasionally nonperiodic bias to the SST, the mean annual SST time series in Fig. 3-1 is considered precise enough for this study.

RESULTS

Fig. 3-1 is a time series of mean annual SST from the region of the Pribilof Islands and Bristol Bay. The annual SST was near the 16-year mean of 4.2 C in 1963 before rising to 5.4 C in 1967. SST then fell to 2.8 C, or 1.4 C below normal, in 1975. Since then, there has been a rapid rise to 5.7 C, or 1.5 C above normal, in 1978. Similar data from Bristol Bay display similar characteristics. Thus, over the five-year period 1967-71, SST decreased by 2.3 C; over the last five years (1974-78) there has been nearly a 3.0 C rise in mean annual SST of the southeastern Bering Sea shelf water. Inspection of the monthly mean SST from which Fig. 3-1 is derived shows that in spring (March and April) 1975 the SST was about 2.7 C below the 16-year mean, while by spring (April and May) 1978 the SST was about 2.0 C above normal, giving a rise of nearly 5.0 C over three years.

It has been suggested that these fluctuations in SST are due to large-scale changes in atmospheric circulation, or more specifically the winter circulation (Niebauer 1980), over the North Pacific and Bering Sea. To outline observations in support of this theory, mean winter 700 mb (3 km) pressure patterns are contrasted with the general winter 700 mb flow patterns (Fig. 3-2) over the period of mean SST illustrated in Fig. 3-1. The mean 700 mb winter flow is generally toward the northeast, parallel to the Aleutian Chain, with relatively weak flow over the Bering Sea.

Fig. 3-3 is an example of upper-air flow from the south bringing warmer air from the Pacific Ocean over the Bering Sea in winter 1966-67. This correlates with, and is probably a primary cause of, the relatively high SST in the mid-1960's (Fig. 3-1).

Fig. 3-4 is an example of a monthly mean 700 mb pressure chart in the winter of 1974-75, showing essentially meridional flow from the arctic, south into the Bering Sea. Some of the flow then turns east and

Figure 3-1. Sea surface temperatures from the Pribilof Island and Bristol Bay regions of the eastern Bering Sea (Niebauer 1978).

flows into the southeast Bering Sea. This cold arctic air appears to cool the underlying Bering Sea (McLain and Favorite 1976). The onset of the decline of SST coincided with anomalous southward penetration of the ice pack (Kukla and Kukla 1974) and had adverse effects on Alaska fisheries (Johnson and Seckel 1976).

A subsequent rise in SST in the middle to late 1970's in the Bering Sea appears to be related, again, to large-scale changes in the atmospheric circulation, which caused southerly flow over the Bering Sea (Niebauer 1978). Fig. 3-5 illustrates the mean 700 mb contours for January 1978, and shows the Aleutian Low over the western Aleutians and strong meridional flow into the southeast Bering Sea from the North Pacific. These patterns are similar to those of the mid-60's and have generally persisted through much of the late 1970's. Namias (1978) has suggested that SST patterns in the North Pacific in November 1976 foretold the strong and persistent air flow from the south over the Bering Sea during winter 1976-77. These air-flow patterns can explain, to a large extent, the high SST in the Bering Sea during this period.

DISCUSSION

That fluctuations in the mean atmospheric circulation are the driving force behind the observed large interannual fluctuations in SST is consistent with the shelf circulation patterns deduced by Coachman and Charnell (1979) and by Reed (1978). Coachman and Charnell point out that the shorter time-scale circulation (~ 25 cm/sec) of this shelf is dominated by tides and wind events, but that the longer time-scale mean flow is weak. Seaward of the 100 m isobath flow is $\sim 2-5$ cm/sec; between the 100 m and 50 m isobaths it is nearly zero. This weak flow on the shelf has the effect of decoupling the transport of mass characteristics, such as heat, from the mean flow (see, for example, Csanady 1976) and makes their horizontal transport a function of large-scale diffusion.

Reed (1978) considered changes in heat content of two $1^\circ \times 1^\circ$ areas inshore of the 100 m isobath. Because of the low net flow in the region, advection of heat was neglected. Reed suggested that gain of heat through horizontal diffusion has little effect on

Figure 3-2. Mean winter (December-February) 700 mb contours (labeled in tens of feet) based on the 26-year period, 1947-72. Flow is generally along the isobars with low pressure to the left (Namias 1978).

Figure 3-3. Mean 700 mb contours (tens of feet) for February 1967 (Posey 1967).

the heat budget and that net radiation is typically the dominant heat flux in the southeast Bering Sea in summer. During the early fall, evaporation becomes important due to increased wind and rapid cooling of the overlying atmosphere. The conclusion drawn from both of these studies is that net heat gain or loss comes about primarily through air-sea interaction, and that due to the low net flow on the shelf, fluctuations in heat are retained longer than in an advective system.

To test this idea, an autocorrelation analysis was done on the deviations from mean monthly SST from the Pribilof Islands for the years 1963-78 (Fig. 3-6). Significant correlation up to 23 months' lag suggests that these anomalies in SST on the eastern Bering Sea shelf do, in fact, persist for at least two years. Thus, if the atmosphere does drive the ocean (Davis (1976)

suggests that it does although Namias (1978) points out that there are probably many and varied feedback loops between ocean and atmosphere and that the subject is highly complex), then Fig. 3-6 suggests that short-term climatic fluctuations may also persist for at least two years. However, Namias and Born (1970) showed that temporal coherence among monthly mean SST patterns in the North Pacific is far greater than any known meteorological coherence; they also found that the SST coherence persists for as long as two years.

That fluctuation in the mean *winter* atmospheric circulation is the driving force behind the observed large year-to-year fluctuations in SST is also consistent with the observation of Coachman and Charnell (1979) that there is significant correlation between June bottom temperatures on the eastern

Figure 3-4. Mean 700 mb contours (tens of feet) for December 1974 (Taubensee 1975).

Bering Sea shelf and the freezing degree days of the previous winter. Moreover, Niebauer (1980) has found significant correlation between mean annual SST (Fig. 3-1) and mean *winter* cloud cover and mean *winter* north-south component of the surface winds but not with *summer* cloud cover or winds.

This is interpreted to mean that the large-scale deviations from mean air and sea temperature in the eastern Bering Sea depend on winter atmospheric advective processes. Insolation is important in the seasonal heating and cooling cycle, but apparently has little effect on interannual deviations from the mean. The strong summer insolation also “caps” the shelf

water between the 50 and 100 m isobaths through the formation of a strong thermocline. This, along with the low net flow, insulates the heat content of this water which has been acquired during the winter. This mechanism probably also accounts for the high SST autocorrelation coefficients up to two years later which, in turn, suggest that short-term climatic fluctuations may also persist for at least two years.

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Figure 3-5. Mean 700 mb contours (tens of feet) for January 1978 (Wagner 1978).

Figure 3-6. Autocorrelation analysis of 1963-79 monthly mean SST from the Pribilof Islands region. The .01 and .05 confidence intervals are indicated.

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Hydrographic Structure Over the Continental Shelf of the Southeastern Bering Sea

4

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ABSTRACT

We synthesize recent work conducted over this exceptionally broad (~ 500 km) shelf which generally has only slow mean flow (≤ 2 cm/sec). Hydrographic structure is little influenced by this flow, but rather is formed primarily by boundary processes: tidal and wind stirring; buoyancy input from insolation, surface cooling, melting, freezing, and river runoff; and lateral exchange with the bordering oceanic water mass. Three distinct hydrographic domains can be defined using vertical structure to supplement temperature and salinity criteria. Inshore of the 50 m isobath, the coastal domain is vertically homogeneous and separated from the adjacent middle domain by a narrow (~ 10 km) front. Between the 50 m and 100 m isobaths, the middle domain tends toward a strongly stratified two-layered structure, and is separated from the adjacent outer domain by a weak front. Between the 100 m isobath and the shelf break (~ 170 m depth), the outer domain has surface and bottom mixed layers above and below a stratified interior. This interior has pronounced finestructure, as oceanic water intrudes shoreward from the weak haline front over the slope, and shelf water (middle domain) intrudes seaward across the 100 m isobath. These domains and their bordering fronts tend to persist through winter, although the absence of positive buoyancy often makes the middle shelf vertically homogeneous.

INTRODUCTION

We selected the title hydrographic "structure" rather than simply "hydrography" because we wish to emphasize the structure, or organization, inherent in the hydrographic distributions. This approach focuses on the shapes of vertical profiles, or rather classes of shapes (e.g., two-layered), rather than on values of temperature and salinity or their correlation (TS diagrams). Thus, we find a large region of the shelf where the temperature and salinity are

vertically homogeneous throughout the year, although the values of temperature and salinity fluctuate over a wide range. We concentrate on the persistent vertical homogeneity and label this region a hydrographic domain. Because vertical profiles control the hydrostatic stability of the water column, and because stability influences vertical mixing, this approach is physically meaningful and useful.

We also concentrate on characteristics of small size, on what can be called the spatial variability. Thus the fronts that separate regions of uniform hydrographic structure (hydrographic domains) are discussed in some detail, as is the finestructure over the outer shelf. One front, for example, has a width of only 10 km and the finestructure has a typical vertical extent of 5 m. It is now possible to resolve such features as fronts and finestructure because samples are taken closer together than formerly.

Our emphasis on hydrographic structure and small spatial scales is not opposed to examination of TS properties or broader spatial scales, but complementary to it. Our description of the shelf hydrographic structure is more meaningful in considering shelf environment from a climatic point of view. We mostly ignore changes at intervals longer than annual, although interannual hydrographic variability is significant (e.g., Overland, Niebauer, and Ingraham, this volume). The major features that we discuss here, however, were observed both in 1976 (the winter of 1975-76 was exceptionally cold, with

extensive ice cover) and in 1977 (the winter of 1976-77 was exceptionally mild, with reduced ice cover). Although altered by interannual changes, the features that we describe persist through these long-term variations.

Because mean flow over the shelf is small, changes in hydrographic properties can be straightforwardly attributed to local processes rather than to advection. For instance, cold temperatures in the lower layer of the middle shelf persist throughout summer (Fig. 4-4). This was once believed to be evidence of mean flow from the northern shelf, but in fact the cold temperatures are caused by local processes: heat loss at the sea surface and complete vertical mixing during winter, followed by the establishment of strong stratification during spring and summer. This stratification insulates the lower layer from downward heat transfer. Especially over the inner two-thirds of the shelf, important characteristics of the hydrography can be explained by local phenomena and advective effects are unimportant.

We complete the introduction by briefly discussing the oceanographic setting, reviewing previous work, and discussing the data. Then we define the hydrographic structure by discussing salient characteristics: domains, fronts, fine structure, winter structure, and river plumes. We then discuss some processes that affect the hydrographic structure: stirring and buoyancy addition, heat and salt transport, and upwelling. Finally we discuss and speculate about aspects of the hydrographic structure.

Setting

The southeastern continental shelf is bordered by the Alaska Peninsula, the Alaska mainland, and a line running southwest from Nunivak Island to the Pribilof Islands and thence following the shelf break southeastward to Unimak Pass. Waters above the shelf receive an annual excess of precipitation over evaporation, as well as freshwater runoff from numerous rivers. Estimating precipitation either from Jacobs (1951) or from station data reported by Brower et al. (1977), and evaporation from Jacobs, a net of about 1 percent of the volume of water over the southeastern shelf is added annually by precipitation minus evaporation.¹ An additional 1 percent is added by river runoff, principally from the Kuskokwim and Kvichak (1,500 to 2,000 m³/sec average discharge from all rivers: Roden 1967, Favorite et al. 1976). In winter, ice covers over 50

¹ Using recent precipitation estimates by Reed and Elliott (1979) would increase this to nearly 2 percent. Reed and Elliott state, however, that their estimates may be inaccurate in the subarctic Pacific.

percent of the shelf, initially appearing inshore in November, often expanding to cover more than 80 percent of the shelf by March, and rapidly disappearing between late April and early June (Favorite et al. 1976, Muench and Ahlnäs, 1976). Ice appears to form near shore and is blown southward during the freezing season (see McNutt and Pease, this volume). Current meter records show that most of the horizontal kinetic energy of the shelf water is tidal: 60-95 percent of the variance in records 9-332 days long was tidal (see Kinder and Schumacher, Chapter 5). Vector mean speeds (≤ 2 cm/sec) were one order of magnitude lower than tidal speeds (~ 20 cm/sec).

Historical review

There has been considerable Japanese, Soviet, and American work done on this shelf. Results of this work have been effectively summarized (Ohtani 1973, Takenouti and Ohtani 1974, Arsenev 1967, Dodimead et al. 1963, and Favorite et al. 1976), and this brief review places more recent results in perspective.

Takenouti and Ohtani (1974) discussed waters above the shelf, which they realized were separated from ocean waters by a "discontinuous zone" (cf. Kinder and Coachman 1978). They further reported that the cold (< 1.0 C) water near the bottom in the middle shelf (cf. Fig. 4-4) was not advected from the Gulf of Anadyr as Kitano (1970) believed, but was formed *in situ* in winter and insulated by strong stratification in summer. Their proposed classification of water masses over the southeastern shelf has been modified by recent findings. Takenouti and Ohtani defined a CW (coastal water) region by its low salinity, but we have found that at the end of winter the salinity may be higher there than in the adjacent convective area (CA—roughly corresponding to our middle domain). Furthermore, the Alaskan Stream (AS) region near the shelf break is misnamed—direct connection with the westward-flowing Alaskan Stream, which exists south of the Alaska Peninsula and Aleutian Islands, is not proved. At the same time, our map of hydrographic domains (Fig. 4-1) is congruent with theirs, and builds upon their insights.

Ohtani (1973) discussed the southeastern shelf in more detail. He mentioned the thermal front that forms between the middle and coastal domains (cf. Fig. 4-6 and Schumacher et al. 1979), and correctly suggested the importance of tidal stirring in forming this front. Ohtani also emphasized vertical stratification in defining shelf water masses, and dwelt less on arbitrary temperature and salinity limits. Again, net inflow of Alaskan Stream water is more tenuous than Ohtani suggested; properties are certainly exchanged

Figure 4-1. Approximate boundaries separating the three shelf (coastal, middle, outer) and the oceanic hydrographic domains. The boundaries are three fronts: inner, middle, and shelf break. These fronts roughly coincide with the 50 m isobath, the 100 m isobath, and the 200 m isobath (shelf break). Profiles from the numbered stations appear in Fig. 4-2.

through the eastern Aleutian passes by vigorous tidal currents, but the net flux of water is not known, and is probably small in any case because of small cross-sectional area (Favorite 1967).²

Arsenev (1967) wrote about water masses and currents of the entire Bering Sea, using many sources, but highlighting Soviet work. He discussed the importance of water mass transformation by fresh-water runoff, insolation, cooling, melting, and freezing. He also recognized the separation of oceanic and shelf waters, but virtually ignored the southeastern shelf in favor of the western shelf, especially the Gulf of Anadyr.

Dodimead et al. (1963) and Favorite et al. (1976) summarized the regional oceanography of the North Pacific, including the Bering Sea. Dodimead et al. (1963) included an appendix on Bristol Bay, and

noted several features that have been elaborated only recently. They reported the inner front that separates the coastal and middle domains as a sharp boundary (cf. Schumacher et al. 1979), and also reported "marked changes" near the shelf break that correspond to the weak haline front there (cf. Kinder and Coachman 1978, Coachman and Charnell 1979). They also noted the patch of cold surface water within Bristol Bay, which they attributed to upwelling. Favorite et al. (1976) showed three domains across the shelf: shelf edge, mid-shelf, and West Alaska Coast (their Fig. 33). Their geographical boundaries nearly coincided with the three domains that we describe in the next section, but they were apparently based on TS relations (cf. Ingraham, this volume). Favorite et al. (1976) also discussed the frontal zone over the slope.

Thus, many of the features that we now recognize as important components of the hydrographic structure of the shelf were reported previously. Among these features are the front over the slope and the inner front farther inshore, the division of shelf waters into distinct domains, and the possible upwell-

² Favorite (personal communication 1979) has pointed out that the distribution of a temperature maximum along the eastern Bering Sea continental slope suggests net inflow to the Bering Sea through passes west of Unimak Pass between 170° W and 172° W.

ing in Bristol Bay. We now know more details and understand these features better, but it is clear that our progress has benefited from these earlier works.

Data

From August 1975 to February 1978, hydrographic casts were made with profiling instruments: STD (salinity, temperature, depth), CTD (conductivity, temperature, depth), or XBT (expendable bathythermograph). Covering all months from February to October, these 1,064 STD and CTD casts are biased towards summer (Table 4-1), but this bias is not a serious limitation because of adequate coverage in February.

The STD and CTD data were calibrated by a water sample, normally taken at the bottom during alternate casts. Calibration temperatures were determined by reversing thermometer, salinity by portable induction salinometer. We claim an accuracy of ± 0.02 C and $\pm 0.02^\circ/\text{oo}$. The XBT profiles were calibrated against nearby CTD casts, and we claim ± 0.1 C. The unusual data processing necessary to examine details of fine structure in vertical profiles of salinity is discussed elsewhere (Coachman and Charnell 1979).

TABLE 4-1

Summary of STD and CTD data

MONTH	NUMBER OF CASTS
February	117
March	65
April	34
May	159
June	213
July	122
August	152
September	184
October	18
TOTAL	1064

These casts were taken during the period August 1975 to February 1978.

HYDROGRAPHIC STRUCTURE

Three hydrographic domains

We have divided the shelf into three structural domains, called the coastal, middle, and outer

domains (Fig. 4-1). These domains are nearly congruent with geographical boundaries previously defined by water masses (e.g., Favorite et al. 1976), and are approximately separated by the 50 m isobath, the 100 m isobath, and the shelf break (close to the 200 m isobath). Our structural domains broaden the criteria previously used for defining the shelf water masses, emphasizing the potential for stratification of the water column (Table 4-2; also cf. Fig 24 in Coachman and Charnell 1979). These domains are most prominent in summer, but are also discernible during the other seasons. Lying seaward of the outer domain, and separated from it by a weak haline front (shelf break front), is the oceanic domain. The oceanic domain completes our scheme, but it is outside the geographic focus of this chapter (Sayles et al. 1979 concentrate on the water overlying the deep basins).

Defining water masses is most useful where temperature and salinity vary slowly at a location (e.g., no surface cooling), or where significant mean advection makes water masses useful in tracing flow. Thus tracing water masses has usefully revealed mean flows in the deep ocean. On this shelf, however, great changes in water mass properties occur annually (Coachman and Charnell 1979, Kinder et al. 1978), and there is little mean flow. Moreover, seemingly reasonable temperature-salinity parameters may prove deceptive. For example, a criterion previously used to describe coastal water has been its low salinity (Takenouti and Ohtani 1974, Favorite et al. 1976), but we now know that during early spring the coastal domain may be more saline than the adjacent middle shelf water (Kinder 1977).

To overcome some of these ambiguities, we have added vertical structure to the criteria. Instead of water mass, we use the word domain, favored by Dodimead et al. (1963), to connote broader criteria than simple temperature-salinity correlations. These domains are geographic entities; energy balances forming vertical structure are closely tied to local geography, so that the domains are also nearly fixed geographically (see *Stirring and buoyancy addition* in the next section of this chapter).

During the summer, vertical structural criteria permit easy separation of the shelf into three domains: homogeneous (coastal), two-layered (middle), and stratified interior (outer; see Table 4-2). These categories are insensitive to particular values of temperature and salinity which vary from year to year (see Niebauer, Chapter 3, for variations of sea surface temperature, and Ingraham, this volume, for interannual hydrographic variations), but depend on the influence of buoyancy input, which

TABLE 4-2
Hydrographic domains in summer

	Coastal	Middle	Outer
Vertical structure	homogeneous	two-layer	surface mixed layer stratified interior bottom mixed layer finestructure
Stratification	very low	very high	moderate
Depth	< 50 m thickness of bottom (tidal) mixed layer	50 m ≤ depth < 100 m thickness of surface + bottom mixed layer	≥ 100 m > surface + bottom mixed layers, thus an interior region exists
Temperature	very warm in late summer (efficient heat transfer throughout water column) (~ 8 to 12 C)	very cold bottom temperature throughout summer (vertical heat transfer impeded by stratification) (~ -1 to 3 C)	moderate (~ 3 to 6 C)
Salinity	generally low (< 31.5 ⁰ /oo), but may be relatively high following winter (> 32 ⁰ /oo: brine drainage during freezing)	moderately low (~ 31.5 ⁰ /oo)	high (> 32 ⁰ /oo)
Influences	river runoff freezing	melting	adjacent water overlying deep basin; Bering Slope Current

This table emphasizes summer conditions, when the domains are most clearly established and when our data are most extensive. These domains remain useful throughout the year: see the section in this chapter on winter structure.

tends to stratify the water column, and tidal stirring, which tends to mix the water column.

An example of each domain from early autumn 1976 (Fig. 4-2) illustrates the three structures. Station 126, in about 50 m of water, is nearly homogeneous in both temperature and salinity, while station 101, in about 70 m of water, is strongly two-layered with vertical temperature and salinity differences of 4 C and 0.4⁰/oo. In still deeper water, station 47 has a surface layer, a bottom layer (not completely mixed), and a stratified interior.

Many stations in the outer domain display strong finestructure in temperature and salinity (Coachman and Charnell 1977, 1979; Kinder 1977; Kinder et al. 1978), but we smoothed the profiles in Fig. 4-2 to emphasize larger-scale features.

A companion view of these domains is shown by plotting vertical temperature differences across the shelf in early autumn 1976 (Fig. 4-3). Shoreward of the 50 m isobath, this difference was generally ≤ 1 C, while between the 50 and 100 m isobaths it commonly exceeded 7 C. Nearer the shelf break,

Figure 4-2. Typical stations from autumn 1976 illustrating the three domains. Coastal (homogeneous), station 126; middle (two-layered), station 101 ; outer (stratified interior), station 47 (see Table 4-2 for domain characteristics). (A) Temperature (°C). (B) Salinity (‰). (C) Temperature-salinity (°C, ‰). Station locations are shown in Fig. 4-1.

Figure 4-3. Maximum vertical temperature difference, surface minus deep ($^{\circ}\text{C}$). The largest differences are in the middle domain, and the smallest in the coastal domain (cf. Fig. 4-1).

intermediate values near 4°C were found. A plot of bottom temperature in autumn 1976 (Fig. 4-4) illustrates the strong insulating effect on the stratification displayed in Fig. 4-3. Even in September, cold temperatures ($<0^{\circ}\text{C}$) remained from the preceding winter, isolated by the very strong two-layered stratification in the middle domain. In contrast, the coastal domain was well mixed and bottom temperatures exceeded 9°C . In the deeper water of the outer domain bottom temperatures were intermediate, generally $3\text{--}6^{\circ}\text{C}$. Obviously, stratification is important to an understanding of the shelf.

Using stratification as an adjunct to water mass analysis is valuable, but Coachman and Charnell (1979) also used the traditional method successfully. They defined a shelf water mass found in the middle domain, and an oceanic ("Alaska Stream/Bering Sea") water mass found above the continental slope (Fig. 4-5). They were then able to explain much of the structure of the outer domain in terms of the lateral mixing along isopycnal surfaces between the shelf and oceanic water masses. The shelf water mass was always less saline and, below 30 m, colder than

the adjacent oceanic water. In spite of annual and interannual variations, there exist throughout the year two water masses, one cold and fresh and the other warm and saline, in juxtaposition along the outer shelf. One important evidence of the interaction of these water masses, finestructure in vertical profiles, is discussed below in a separate section.

A combination of categorizing by vertical structure and by traditional water mass techniques is more useful than either used separately. For examining the shelf alone, structural categories are most distinct, but for understanding interaction with waters overlying the deep basin, water mass analysis is useful. As Coachman and Charnell (1979) implied, these two views are interdependent.

Fronts

The four hydrographic domains (coastal, middle, outer, and oceanic) are separated by three fronts. The front separating the coastal and middle domain is much narrower than the domains themselves, and so is legitimately called a front. The other two transitions are much broader relative to their adjoining

Figure 4-4. Bottom temperature ($^{\circ}\text{C}$), late September and early October 1976. Even in autumn, low temperatures persist in the bottom layer of the middle domain.

domains, but since they have been called fronts (e.g., by Iverson et al. 1980), we adopt this usage also. Proceeding seaward, we label these fronts as inner, middle, and shelf break (Fig. 4-1).

The inner front separates the homogeneous coastal domain from the two-layered middle domain. It was hinted at by Dodimead et al. (1963), and noted by Muench (1976) farther north. Schumacher et al. (1979) have called it a structural front, to stress the separation of two vertical structures or marked change of stratification rather than the separation of two water masses. The front is about 10 km wide and generally follows the 50 m isobath (Fig. 4-1). Approaching shallower water from the middle domain, isotherms, isohalines, and isopycnals all spread from the thin thermocline, halocline, and pycnocline over the middle shelf (Fig. 4-6). Within 10 km the vertical hydrographic structure changes from distinctly two-layered to nearly homogeneous. Away from the front, within the strongly stratified middle domain, the thickness of the bottom mixed layer (as judged by nearly isothermal and isohaline profiles) is

nearly 50 m, about the same as the total water depth where the front is found. In general, we find that over the middle shelf the bottom mixed layer is ~50 m thick, the surface mixed layer is 15-20 m thick, and the front occurs where the water depth approximately equals the thickness of the bottom mixed layer (i.e., 50 m); the strongest stratification occurs where the sum of the bottom and surface mixed layer thickness equals the water depth (i.e., 20 m + 50 m = 70 m).

During winter, the middle and coastal domains are nearly vertically homogeneous following surface cooling, freezing, and vigorous wind stirring in fall and winter (but see our section on winter structure for an important exception). Frontogenesis occurs with the addition of meltwater during the ice breakup in spring. As this initial stratification forms, it is reinforced by insolation, so that later in summer thermal stratification is primarily responsible for vertical density differences. Schumacher et al. (1979), Kinder (1977), and Kinder et al. (1978) reported details of this front, and Simpson and

Figure 4-5. Temperature-salinity correlations, middle domain (SHELF) and oceanic domain (ALASKA STR./BERING). Envelopes drawn from data gathered in August 1976 and May 1977 illustrate the warmer and saltier oceanic water at the same density as the cooler and fresher shelf water, and interleaving occurs across the outer domain. (From Coachman and Charnell 1979.)

Pingree (1978) summarized features of similar fronts over the European continental shelf.

The shelf break and middle fronts are less clearly describable. Overlying the continental slope, the shelf break front separates the oceanic domain from the outer domain, but the width of this front is similar to that of the outer domain. Similarly, the middle front which divides the middle and outer domains near the 100 m isobath (Fig. 4-1) is broad

and ill-defined compared to the inner front. Nevertheless, the shelf break and middle fronts are both real and important components of the hydrographic structure.

Kinder and Coachman (1978) described the shelf break front and recognized its essentially haline character. The front is revealed by a change in the horizontal salinity gradient (from nearly zero over the deep basin to about 4×10^{-3} g/kg/km over the outer

Figure 4-6. The structural (inner) front separating the coastal and middle domains. This line was between Nunivak Island and the Pribilof Islands. (A, B) Temperature and salinity cross sections, and (C) sequential temperature profiles from June 1976. The sections are based on stations separated by about 10 km. (D, E) The same section based on widely spaced CTD stations in autumn 1976. (From Schumacher et al. 1979.)

shelf), by isopycnals extending from the shelf to intersect the sea surface above the slope, and by isolines downwarped beneath the front. Available winter data show that this front persists throughout the year.

Coachman and Charnell (1979) and Coachman (1978) examined this region in more detail, and described this transition zone as two broad fronts, one over the slope and one farther inshore near the 100 m isobath. Between these two transitions, each of which has a large horizontal salinity gradient ($\sim 10 \times 10^{-3}$ g/kg/km), is a region of very small gradient (Fig. 4-7). The transition near the shelf break corresponds to the front described by Kinder and Coachman (1978), while the inner transition corresponds to the middle front separating the middle and outer domains (Fig. 4-1).

At different times when examined by different distributions, these broad transitions do appear truly frontal. For instance Coachman and Charnell (1979) showed a mean cross-shelf temperature section for August 1976 that clearly showed a thermal front near the 100 m isobath, and Coachman (1978) showed strong evidence of a front delineated by particulate

and chlorophyll *a* concentrations in April 1978. Over the slope Kinder and Coachman (1978) showed a shallow weak-density front in an August 1972 section, and we show a weak-salinity front from February 1978 (Fig. 4-9). Kinder and Coachman (1978) also showed large dissolved phosphate and nitrate gradients across the shelf break front in July 1974. Both the middle and shelf break fronts generally appear broad and therefore weak, but occasionally they are manifest as sharp fronts in various properties. The shelf break front, however, can always be detected as a weak front in salinity.

Finestructure and density inversions

Finestructure, the layering of vertical profiles on scales from 1 to 25 m (Fig. 4-8), is a salient feature of the outer domain (Table 4-2). The distribution of the finestructure and the mixing physics associated with it are clues to understanding cross-shelf fluxes.

Horizontal distributions of the occurrence of finestructure over the shelf showed that it was common between the shelf break and the 100 m isobath, and occurred elsewhere only rarely. During

Figure 4-7. Vertically (0-100 m) and horizontally (along-isobath) averaged sections across the shelf from May 1976, August 1976, and May 1977. Transitions in the salinity gradient mark the 100 m isobath (middle-outer domains; middle front) and the shelf break (outer-oceanic domains; shelf break front). (From Coachman and Charnell 1979.)

Figure 4-8. A superb example of temperature and salinity finestructure in August 1976. Finestructure, although often less pronounced than this, was present at most outer domain stations. (From Coachman and Charnell 1979.)

1976, a year when the shelf was surveyed extensively, finestructure in the outer domain was reported in March (Coachman and Charnell 1977), in June (Kinder 1977), in August (Coachman and Charnell 1979), and in September-October (Kinder et al. 1978). Only a few stations with finestructure were reported outside the outer domain (e.g., Kinder 1977, Fig. 22), and data from 1977 and 1978 also conform to these distributions. As Coachman and Charnell (1979) discussed, the finestructure occurs in the interior of the water column, below the surface mixed layer and above the bottom mixed layer.

Within this interior region, warmer and saltier oceanic water intrudes shoreward while cooler and fresher shelf water intrudes seaward. As interpreted by Coachman and Charnell (1977, 1979), the outer domain is a region of lateral (i.e., cross-flow, and here also cross-isobath) water mass interaction with interleaving of water masses occurring at finestructure scales. Such interleaving, when water masses of similar density but differing temperature and salinity values mutually intrude, has been observed in many other locations (e.g., see *J. Geophys. Res.* 83(C6) 1978). Occurrence of finestructure throughout the

outer domain, best documented in 1976 (a year with extensive ice cover and late ice breakup) but also observed in 1977 and 1978, implies that finestructure is an inherent part of mixing across the outer domain. An essential stage in mixing large masses of water is the reduction of the spatial scale of identifiable water parcels, until a scale is eventually reached at which molecular diffusion is effective. As the spatial scales decrease, spatial gradients increase, as does the surface area of the boundary, and so mixing progresses. Interleaving on finestructure scales is the initial scale reduction. While finestructure features are only a few meters in vertical extent, they apparently extend horizontally for tens of kilometers. In both March and August 1976, Coachman and Charnell (1977, 1979) traced temperature-salinity correlations within layers for distances of about 100 km.

One startling result of Coachman and Charnell's work was the discovery of a static instability in a layer about 10 m thick in March 1976 and many smaller-scale instabilities of a few meters' thickness in summer. The larger instability was clearly resolved by the instrumentation used (standard CTD vertical profiling system), and had an apparent lifetime of about one week. They speculated that it was formed by interaction between strong winds and the seasonal ice cover. The smaller instabilities were poorly resolved by the standard CTD profilers used, but Postmentier and Houghton (1978) measured nearly identical features over the oceanographically similar slope region south of New England using a higher-resolution profiler. Both Coachman and Charnell (1979) and Postmentier and Houghton (1978) invoked differential diffusion of temperature and salt to explain the smaller instabilities. Because heat diffuses more rapidly than salt at molecular scales (it is easier to transfer energy than mass), adjacent layers of water can become convectively unstable on small scales, either through salt fingers (warm and saline water overlying cool and fresh water) or through double diffusion (cool and fresh water above warm and saline water). In the outer domain, the conditions for salt fingers exist at the lower interface of shoreward-intruding basin water, while the conditions for double diffusion exist at the upper interface.

Winter structure

The discussion of hydrographic domains and fronts mostly reflects summer conditions, but winter conditions are more interesting than might be expected. In winter, waters above most of the shelf usually are vertically homogeneous, with two exceptions. In the outer domain warmer (but more saline

and therefore denser) water from the oceanic domain intrudes beneath cooler and fresher shelf water, thus maintaining stratification. Elsewhere, low-salinity water from melting ice may stratify water that was well mixed during autumn and winter (by wind stirring and surface cooling).

A cross section taken from southeast of the Pribilofs toward Cape Newenham in February 1978 (Fig. 4-9) illustrated intrusion of the basin water. Between the shelf break and the 100 m isobath (i.e., outer domain) water warmer than 3.5 C and saltier than 32.5‰ intruded beneath shelf water which was both colder and fresher. Inshore of the 100 m isobath the water column was well mixed, colder (<2.5 C) and fresher (<32.25‰). Data from several stations with similar profiles, saltier and warmer near the bottom, were also taken near the Pribilof Islands during April and May 1978, and Coachman and Charnell (1977) showed data with this character taken in March 1976. There is sufficient coverage of the outer domain during late winter and early spring to suggest that cold and fresh shelf water overlies warmer and more saline basin water, and that this domain remains stratified during winter.

Figure 4-9. Temperature (°C) and salinity (‰) across the shelf in February 1978. This section is from southeast of the Pribilofs toward Cape Newenham. In the outer domain the deeper water is warmer, but more saline and therefore denser, than the shallower water.

Melting ice in the middle shelf can also cause stratification during the winter, but inshore within the coastal domain mechanical stirring keeps the water column well mixed. In February 1978 we observed (by satellite imagery, see Fig. 5-11, Chapter

5) that ice near Nunivak Island moved about 100 km southeast, into an area previously free of ice. About ten days later we measured hydrographic properties near this ice, which was melting. Away from the ice (~20 km), sea surface temperatures were near 0 C, and temperature profiles were vertically homogeneous (Fig. 4-10). Within the ice (where water depths exceeded 50 m), however, the water column was stratified in two layers. In the shallow layer temperatures were near freezing (~-1.73 C) and the salinity was lower than in the homogeneous water.

Figure 4-10. Temperature (°C), salinity (‰), and density (kg/m^3) profiles near the ice edge in February 1978. Dashed profiles were typical away (>20 km) from the ice or where water depth was less than about 50 m. Solid profiles were typical near the ice where water depth exceeded 50 m.

Below the weak pycnocline, salinity and temperature were similar to values away from the ice. The decrease of temperature and salinity probably resulted from ice melting, about 30 cm of ice for Fig. 4-10. The transition between two-layered and homogeneous conditions occurred near the 50 m isobath, as in summer. Inshore of the 50 m isobath, the water column was homogeneous with or without ice.

We do not know how persistent this winter stratification is, but we suspect that the weak stratification found in water deeper than 50 m was fragile, dependent in part on the continued presence of ice. Once the upper layer cools to the freezing point, ice stops melting. As stirring erodes the pycnocline, however, heat remaining in the bottom layer is mixed upward, presumably melting ice and adding light meltwater. The continued presence of ice above a stratified water column in winter apparently favors continued stratification, suppressing wind stirring, limiting surface

heat loss, and maintaining a reservoir of potential meltwater. The question of whether ice cover generally affects the hydrographic structure over the middle shelf in winter and to what extent this structure in turn influences the resulting stratification during the ice-free season remains unanswered, but the ice may be important through the following summer. It is clear that the eventual melting of ice in spring is important in stratifying the middle shelf domain (Schumacher et al. 1979).

River plumes

The local effects of river discharge have received little attention because most oceanographic data have been collected away from the coast. Satellite images and sparse hydrographic data suggest that river plumes (defined, say, by salinity lower than 25‰) remain near shore, flowing anticlockwise around Bristol Bay, and leaving the southeastern shelf to the north (much of this water may flow through Etolin Strait, inshore of Nunivak Island). Large discharges of fresh water can stratify the water column in the coastal domain, and may form fronts (see Garvine and Monk 1974 for a description of the frontal plume of the Connecticut River in Long Island Sound).

Straty (1977) reported observations made in Bristol Bay during 1966. He traced the anticlockwise nearshore flow of river water around Bristol Bay using dye, drift cards, and salinity measurements. He reported no fronts associated with the Naknek, Kvichak, Egegik, and Ugashik rivers, probably because of vigorous tidal stirring in the shallow (less than 20 m) bay. Conditions may be similar in Kuskokwim Bay farther west, but we have no data there. The direct effects of the river discharges appear to remain within a few tens of kilometers of the coast, providing an inshore boundary of the coastal domain.

PROCESSES THAT AFFECT THE HYDROGRAPHIC STRUCTURE

Many processes can form, alter, or erase features of the hydrographic structure. We have grouped such processes into three somewhat arbitrary categories. First we discuss the addition of heat and salt and their transport across the shelf, processes which directly transform water masses. Next we focus on the interplay of mechanical stirring and buoyancy addition. These processes determine the stratification, a key element of the structure and a strong influence on the transport. Finally, we speculate on the possibility of upwelling, which may affect the hydrographic distributions in Bristol Bay.

Heat and salt transport

Transformations of water masses on this shelf occur locally through the addition of heat and salt. Because of cooling and heating at the surface, evaporation and precipitation, and freezing and melting, relatively large fluxes of heat and salt occur at the sea surface. To a lesser extent horizontal mixing and river runoff at lateral boundaries influence temperature and salinity. Because of the low mean flow on the shelf, and because of the shelf's great width, water mass properties are more likely to result from local phenomena (e.g., insolation and melting) than from advection. Great changes occur annually in the flux of heat and salt at shelf boundaries, so that water masses vary annually also.

In the middle and coastal domains the change in heat content of the water is primarily balanced by heat transfer through the sea surface; horizontal advection and horizontal turbulent diffusion are much smaller. Reed (1978) calculated a heat budget for an area (1° lat. $\times$ 2° long.) of the middle domain for summer 1976. The local rate of heat change was balanced over the summer by net surface exchange within 10 percent (excellent agreement for such budgets). During the summer, most of this surface exchange was radiative, but by early fall evaporation was important. Over the fall, winter, and early spring, the terms incorporating phase changes (evaporation, freezing, and melting) share importance with radiation terms. The net surface exchange retains its importance, however: Coachman and Charnell (1979) found a high correlation ($r = -0.96$, $n = 12$) between mean lower-layer temperatures in June over the middle shelf and degree-days of frost for the preceding winters. Reed's (1978) results can probably be extrapolated into the coastal domain also. Although the vertical heat distribution differs there (Fig. 4-2), the horizontal terms and surface exchanges are probably similar.

In the outer domain, however, horizontal terms apparently are more significant. Since mean flow is 2-10 cm/sec, advection cannot be ignored, and lateral exchange, as evidenced by finestructure, may be even more important. The strong annual cycle which Coachman and Charnell (1979) showed for this region—approximately the seaward half of the shelf waters and those over the slope—was caused by surface exchange. Below the surface mixed layer, however, they showed large-amplitude finestructure (Coachman and Charnell 1977, 1979), with warmer shoreward intrusions originating in the oceanic domain and colder seaward intrusions originating over the shelf (Fig. 4-8). These lateral interleavings are strong evidence of lateral exchanges of heat and salt,

with the shelf water (colder and fresher) gaining heat and salt.

Various attempts have been made to estimate the horizontal heat flux in terms of a bulk heat conductivity such that the turbulent horizontal heat flux is given by:

$$\rho C_p K_h \frac{\partial T}{\partial X} \quad (\text{J m}^{-2} / \text{sec})$$

$$\rho = \text{density of water} \quad (\text{kg/m}^3),$$

$$C_p = \text{heat capacity} \quad (\text{J/kg/}^\circ\text{C}),$$

$$\frac{\partial T}{\partial X} = \text{horizontal temperature gradient} \quad (^\circ\text{C/m}),$$

and

$$K_h = \text{horizontal conductivity} \quad (\text{m}^2 / \text{sec} = 10^4 \text{ cm}^2 / \text{sec}).$$

This is sometimes a poor approximation of the physical processes (which are hidden within K_h), and K_h is often not constant. Nevertheless, such estimates remain useful for modeling and estimating cross-shelf fluxes. Kinder et al. (1978) calculated a heat balance for the lower layer of the middle shelf over summer 1976 and estimated $K_h \sim 1.7 \times 10^6 \text{ cm}^2 / \text{sec}$ ($= 1.7 \times 10^2 \text{ m}^2 / \text{sec}$). They similarly estimated vertical conductivities in the middle shelf, and values ranged from $7 \times 10^{-3} \text{ cm}^2 / \text{sec}$ to $5 \times 10^{-1} \text{ cm}^2 / \text{sec}$. Because of the strong stratification, the lowest values approached molecular diffusion ($\sim 1.4 \times 10^{-3} \text{ cm}^2 / \text{sec}$). These estimates by Kinder et al. (1978) were probably maxima, since they assumed that all local change had been caused by one-dimensional diffusion, either vertical or horizontal. Coachman and Charnell (1979), applying a model proposed by Joyce (1977), estimated $2.8 \times 10^6 \text{ cm}^2 / \text{sec}$ and $1 \times 10^6 \text{ cm}^2 / \text{sec}$ for the middle and shelf break fronts and $20 \times 10^6 \text{ cm}^2 / \text{sec}$ for the outer domain between fronts.

Kinder and Coachman (1978) calculated a salt budget for the entire shelf. Since fresh water is added annually at the coast by river runoff, and because precipitation exceeds evaporation over the Bering Sea, there must be a flux of salt shoreward to maintain the long-term mean salinity distribution. For the shelf as a whole, the largest term (>99 percent of salt flux) is advection: relatively saline water from the oceanic domain flows onto the western shelf to supply the Bering Strait outflow ($\sim 1.0 \times 10^6 \text{ m}^3 / \text{sec}$). Over the southeastern shelf, however, the salt balance is not advective (because of low mean flow). Kinder and Coachman (1978) estimated a diffusivity of $3 \times 10^6 \text{ cm}^2 / \text{sec}$ for the cross-shelf salt

flux. Calculating diffusivity for the middle domain over summer 1976, Kinder et al. (1978) obtained $1.1 \times 10^6 \text{ cm}^2 / \text{sec}$, using the same method as for thermal conductivity (the salt diffusion equation was analogous to that of heat).

Kinder and Coachman (1978) suggested that the cross-shelf salt flux was driven by the tides, as a "tidal diffusion," because most (~ 90 percent) of the kinetic energy over the shelf is tidal (Chapter 5, this volume). The tidal current, if appropriately correlated with salinity variations over the tidal cycle, could cause a significant flux of salt across the shelf. Coachman and Charnell (1979) showed, however, that in the outer domain lateral interleaving on vertical fine-structure scales is ubiquitous and represents cross-shelf mixing. Tidal diffusion still remains tenable for the middle and coastal domains, and the tides do contribute most of the turbulent energy (via the bottom frictional layer and velocity shear) within the outer domain.

Kinder et al. (1978) also reported another means of salt flux—ice transport. During the ice-covered part of the year, satellite imagery often shows open water south of east-west zonal coasts: south of St. Lawrence Island, south of Nunivak Island, and northern Kuskokwim Bay are typical examples (see Muench and Ahls 1976 and chapters by McNutt and Pease, this volume). During the spring of 1976 (Kinder 1977) and to a lesser extent in 1977, water with elevated salinities ($> 32.5^\circ / \text{oo}$ in June 1976) was found in Kuskokwim Bay. Our explanation is that ice freezing in Kuskokwim Bay is blown seaward, leaving behind the brine that drains during freezing. We do not know accurately the amount of ice exported from the coastal domain annually in this way, nor do we know the salinity of the exported ice. Kinder et al. (1978) estimated that this divergence of ice transport may account for a mean salt flux of 6 t/sec ($1 \text{ t} = 10^6 \text{ g}$) shoreward from the middle to the coastal domains. This is about 10 percent of the mean salt flux (50 t/sec) estimated by Kinder and Coachman (1978) for the southeastern shelf. As Coachman (1979) pointed out, this mechanism may be generally important at high latitudes; hypersaline water relict from the previous winter has been found not only in Kuskokwim Bay, but recently in Norton Sound (Chapter 6, this volume), Kotzebue Sound (Kinder et al. 1977), and lagoons adjoining the Beaufort Sea (Wiseman 1979). Since this mechanism causes a net freshening of the middle domain, it may partly explain the correlation that Coachman and Charnell (1979) reported between yearly mean temperatures and yearly mean salinities over the middle shelf: both cooling of the middle shelf

waters and the export of ice from the coastal to the middle shelf domains may be causally related to severe (cold and windy) winters, when southward outbreaks of cold and dry continental air cause more ice formation (see Overland, Chapter 2, this volume).

Stirring and buoyancy addition

A water column is stably stratified if the density increases towards the bottom. During spring and summer, this stable water column usually prevails because lighter, more buoyant fluid is added at the surface (ice melting, precipitation, or freshwater runoff) or because the surface waters become less dense as a result of warming (insolation). Alternatively, the addition of dense water at the bottom (intrusion of oceanic-regime water onto the shelf) makes the surface waters more buoyant than the bottom waters (Fig. 4-9). Processes that tend to stratify the water column stably by decreasing the density of the near-surface, we call positive buoyancy additions. If dense water is added at the surface (brine drainage during freezing) or if the surface waters are made denser (radiative cooling or evaporation), then the water column becomes less stratified or less stable. If water becomes denser than that below it, then the water column is unstable and vertical mixing (overturn) occurs. We call processes that destabilize the water column negative buoyancy additions. By buoyancy addition we mean any change in water properties that alters the mean density of the water column (as distinguished from mechanical stirring that redistributes the density).

Mechanical stirring is an important process tending to mix the water column. Over this shelf, the main source of stirring is the tidal currents and a secondary source is the wind (Schumacher et al. 1979, Simpson and Pingree 1978). Most tidal stirring power (turbulence) is generated near the bottom, most wind stirring power (turbulence) at the surface. Thus, we attribute the surface mixed layer to wind stirring, and the bottom mixed layer to tidal stirring. Station 101 in Fig. 4-2 illustrates these two layers in which mechanical stirring is sufficient to keep temperature and salinity homogeneous over layers of 20 m or more thickness. At Station 126, stirring had overcome any stabilizing effects of positive buoyancy addition, and the entire (50 m) water column was well mixed (neutrally stable).

Another way of viewing these two tendencies is to consider the potential energy of two water columns. Consider the first, like Station 101, to consist of two layers, while the second, like Station 126, is completely mixed. If both columns have the same

vertically averaged temperature, salinity, and thus density (assuming a linear equation of state), then all points from both stations fall on the same straight line on the TS plane; it is the vertical structure that differentiates between the two distributions. It requires mechanical work to mix the two-layered water column so that it looks like the homogeneous water column, because the center of mass of the well-mixed water column is higher than in the two-layer column. Over the Bering Sea shelf the primary source of this mixing energy is the tides.

When the cross section across the inner front was made in 1976 (Fig. 4-6) we found two water columns like those just described. The homogeneous water column on the coastal domain side of the front could have been made by completely mixing the water column on the middle domain side of the front. On the shoreward side of the front, the tidal stirring was just competent to overcome the buoyancy addition from melting ice and insolation; thus fresh water and heat were mixed throughout the water column. On the seaward side of the front, however, stirring was inadequate. A wind-stirred surface layer and a tidally stirred bottom layer met at a sharp pycnocline.

This interplay of buoyancy addition and stirring has some positive feedback. Stratification suppresses vertical mixing so that mixing is impeded after stratification forms, and as further buoyancy is added stratification increases. This added stratification further suppresses mixing, and so forth. This feedback helps explain why the transition between the coastal and middle domains is so sharp: stratification enhances stratification, and well-mixed structure likewise tends to persist. Over the middle shelf the surface mixed layer and the bottom mixed layers meet, making the vertical structure distinctly two-layered. The middle front, separating the middle and outer domains, marks the limit of the ability of the two homogeneous layers to encompass the entire water column. Seaward of this front, in the outer domain, an interior region exists between the two mixed layers. Finestructure exists only in this interior; elsewhere it would be vertically mixed by stirring. Table 4-3 emphasizes the roles of stirring and buoyancy addition in forming the hydrographic domains.

We can get a feeling for the reason why the domains are closely tied to the isobaths (50 m, 100 m, shelf break) by following the formalism of Simpson and Hunter (1974), who examined a front like our inner front near the British Isles. They compared the rates of addition of potential energy by insolation and by stirring.

For a two-layered water column, potential energy (V) addition rate is approximately:

$$\frac{dV}{dt} = \frac{\alpha Qgh}{2C\rho} \quad (\text{J m}^{-2}/\text{sec}) \quad (\text{J} = \text{Joule})$$

α : a thermal expansion coefficient
($\text{kg}^\circ\text{C}^{-1} \text{m}^{-3}$)

Q : insolation ($\text{J m}^{-2}/\text{sec}$)

g : acceleration of gravity (m/sec^2)

h : water depth (m)

C : specific heat ($\text{J/kg}/^\circ\text{C}$)

ρ : density (kg/m^3)

The major annual change is in the insolation term (Q); other buoyancy terms (e.g., melting ice) could be added easily.

The turbulent energy available for stirring is simply:

$$\frac{dE}{dt} = k \rho U^3 \quad (\text{J m}^{-2}/\text{sec})$$

where: k = drag coefficient, ρ = density (kg/m^3), and U^3 = mean of cubed speed (m^3/sec^3).

Most of this power (note that $1 \text{ J/sec} = 1 \text{ watt}$) does not mix the water, but the relative amount (1 percent or so) that does go into mixing seems constant for a given flow regime (e.g., near the inner front).

We can see that the buoyancy addition term has small changes across the shelf in all of its terms but h , while in the stirring term U^3 changes across the shelf. The tidal current, U , is also a function of depth (h) and position on the shelf (Pearson et al., Chapter 8, this volume), so that both buoyancy addition and stirring are dependent on location. Although neither buoyancy addition nor stirring changes very rapidly at a given location, the tidal currents vary significantly over two-week cycles (fortnightly tide), winds vary, and buoyancy input changes diurnally and annually; but an important variation across the shelf can be seen by taking the ratio of dE/dt and dV/dt . The result is a constant times U^3/h ; across the shelf, from the outer to the coastal domain, this changing ratio reflects the changing balance between buoyancy and stirring. Nearshore, since U^3 is large and h small, stirring prevails. Farther offshore, because U^3 decreases and h increases, stratification (given sufficient Q or other buoyancy source) prevails. We even found that this held in February 1978 when we took measurements in melting ice: seaward of the 50 m isobath the water column was

TABLE 4-3

Stirring and buoyancy addition in the hydrographic domains

Coastal domain

Throughout the year tidal and wind stirring produce adequate mixing power to overcome normal sources of buoyancy: insolation, melting ice, and river runoff. Exceptions to this are probably short lived, except in river plumes within 10-20 km of the coast. (Water depth < thickness of tidal-mixed layer.)

Middle domain

Tidal and wind stirring are inadequate to mix the entire water column during the high buoyancy-addition season (spring and summer). The vertical structure during that season is two-layered: a wind-stirred surface layer and a tidal-stirred bottom layer separated by a sharp pycnocline. During fall and winter, when buoyancy addition is usually negative, the vertical structure is uniform, but the potential for stratification remains. Melting ice can establish two-layered stratification, even in winter. (Water depth = thickness of tidal-mixed layer + thickness of wind-mixed layer.)

Outer domain

The surface mixed layers and bottom mixed layers do not meet; the pycnocline is weaker than in the middle domain and there is an interior region separating the mixed layers. Fine-structure is ubiquitous within this stratified interior region. Even in winter, negative buoyancy and stirring are insufficient to mix the water column completely. More saline water from the oceanic regime makes the deep column denser than the surface waters, even if the surface waters are cooled to the freezing point. (Water depth > thickness of tidal-mixed layer + thickness of wind-mixed layer.)

See also our Table 4-2 and Figure 24 in Coachman and Charnell (1979).

two-layered, while shoreward of the 50 m isobath the column was well mixed (the 50 m isobath coincides with the inner front during summer). Thus the potential for stratification (expressed by U^3/h), is always present, requiring only sufficient buoyancy addition to be realized.

Our data do not reveal variability in frontal location, either on short time scales such as diurnal

or fortnightly, or longer scales such as interannual. There is undoubtedly some variation in the location of fronts on many scales, but the inner front is tied closely to its mean position by the variation of U^3/h , and similar considerations probably affect the middle front's position also. Stirring and buoyancy addition form the vertical hydrographic structure within the coastal and middle domains, and modify the structure of the outer domain.

Upwelling

The cold surface patch observed in Bristol Bay during summer has been ascribed to upwelling. Myers (1976) documented the frequent occurrence of cooler surface water in Bristol Bay during spring and summer, and our own data also showed this (Kinder et al. 1978). Myers attributed this to upwelling forced by an Ekman convergence in the bottom boundary layer. This convergence was caused by a mean cyclonic flow that approximately follows the 50 m isobath.

Arguments based on hydrographic evidence from 1969-70 presented by Myers favored upwelling of water originating southwest of the cool surface patch rather than local vertical mixing, but the explanation of this upwelling was incomplete. The mean flow is only about 2 cm/sec, while tidal speeds are about 20 cm/sec (Chapter 5, this volume). Thus, the tidal kinetic energy is 100 times that of the mean flow, and tidal effects may be more important than the mean flow. For the Ekman convergence to work, water must be forced upward against stratification, rather than forced horizontally to the west or southwest (where there is no mean flow). Moreover, Myers hydrographically inferred that the source of upwelled water is southwest of Bristol Bay, but his proposed Ekman convergence requires a source to the east and north. A possible alternative, strong wind during summer, occurs too infrequently to account for this persistent feature. In the open ocean, with upper layers moving faster than lower layers, large (nearly geostrophic) cyclonic features are associated with isopycnals that dome upward; perhaps the mean flow does influence the observed distributions in Bristol Bay. A secondary circulation would then be necessary to maintain the density structure against tidal stirring and mixing. On the other hand, a combination of vertical mixing driven by tidal currents and freshening of inshore waters by river runoff could produce the observed hydrographic distributions. This seems more in harmony with processes over the remainder of the shelf, but is no more proved than the upwelling hypothesis.

In summary, cool surface water appears often in Bristol Bay during spring and summer, and Myers (1976) presented hydrographic evidence that this results from upwelling. Whether this persistent feature actually results from upwelling, from another dynamic response to the flow regime, or from vertical mixing, however, is not known.

DISCUSSION

Cross-shelf fluxes

On the basis of conservation of heat and salt, we have discussed some estimates of cross-shelf fluxes in terms of diffusion coefficients. Knowledge of these fluxes, and of the mechanisms driving them, is important for both conservative and non-conservative material, e.g., larvae, nutrients, plankton, and petroleum. Especially in summer, dispersion characteristics differ in the different domains. Vertical exchanges are severely damped in the strongly stratified middle domain, while complete vertical mixing occurs rapidly in the tidally stirred coastal domain. Dispersion also differs horizontally in the three domains. In the outer domain interleaving on fine-structure scales is an important component of mixing processes, but no finestructure is found in the inner two domains. Mixing, although probably driven by the dominant tidal currents in both nearshore domains, differs between the middle and coastal domains. Over the two-layered middle shelf horizontal transport may differ markedly in each layer (e.g., a nearly estuarine two-layer flow), but this is unlikely in the vertically homogeneous coastal domain.

There is also a question of steadiness of these fluxes: how much do they vary and over what time scales? Coachman and Charnell (1979) estimated that the horizontal salt flux in the outer domain was three to four times greater than the fluxes at the shelf break and middle fronts. This implies a divergence (depletion) of salt transport near the shelf break and a convergence (accumulation) near the 100 m isobath—an imbalance which cannot persist over long periods without altering the observed long-term salinity distribution. There is some annual variation in fluxes, as the hydrographic structure, wind stress, and ice cover all change significantly over the year. The variability of these fluxes, and particularly their timing (or phasing) with respect to critical biological events, may be more important than the mean fluxes. As yet, we can only roughly estimate mean values, and we do not understand the processes that drive these fluxes.

Fronts

It is not clear whether the fronts separating the hydrographic domains are convergences or divergences—whether they enhance or inhibit mixing. These boundaries separate distinct hydrographic domains and probably dynamic ones, and they have large gradients in various properties. The methods of transport of passive properties, such as salt and nutrients, and of dynamic properties, such as momentum and vorticity, probably change across these fronts, and these changes are most clearly seen in the vertical hydrographic structure. Intuition suggests that fronts are convergences (James 1978), and that cross-frontal exchanges are impeded (e.g., methane distributions shown by Cline, this volume). A convergence throughout the depth and length of the inner front, for instance, seems unlikely; but neither observation nor modeling have answered the questions of convergence and cross-frontal mixing.

There is evidence of year-to-year (Coachman and Charnell 1979) and annual (Schumacher et al. 1979, Kinder and Coachman 1978, Coachman and Charnell 1979) variability of the fronts, and Schumacher et al. (1979) reported wavy features in satellite images of the inner front that imply more rapid variability. The longer-term fluctuations seem related to changes in atmospheric forcing (e.g., insolation, temperature, storms), and the wavy features may be frontal instability (inherent). Further understanding of these changes will probably add knowledge of variations in cross-shelf fluxes.

Role of ice

Ice, with strong annual and interannual variation, influences the hydrography of this shelf in several different ways. These effects are both local and shelf-wide.

Locally, ice affects the energy balance and vertical distributions of heat and salt. Ice cover effectively insulates the water and slows heat transfer (both radiative and conductive-convective), and ice covered with snow has high albedo, reflecting incoming short-wave radiation. Freezing and melting also alter the distribution of heat and salt. Ice acts as a buffer for temperature as changing heat balance alters the amount of ice present at nearly constant temperature. Local freezing and then melting causes a vertical redistribution of salt, so that a water column that has uniform salinity in fall may have haline stratification in spring.

Freezing and melting also influence shelf-wide distributions of heat and salt. Freezing nearshore and melting offshore transport both salt and heat shore-

ward. Waters offshore are cooled directly not only by the atmosphere, but by melting ice that originally formed nearer shore. Because these processes are forced by weather, changes in the winter weather are manifest in ice cover and therefore in the shelf hydrography.

Ice processes thus affect the shelf hydrographically in two ways: through melting and freezing, ice locally redistributes salt and heat in the water column, changing the vertical stratification; and it directly influences shelf-wide heat and salt budgets by acting as an insulating cover while transporting salt and heat.

SUMMARY

The southeastern shelf has a distinct hydrographic structure. Proceeding seaward from the coast in summer one encounters the vertically homogeneous coastal domain, the inner (structural) front, the two-layered middle domain, the middle front, the outer domain, the shelf break front, and finally the bordering oceanic domain (Fig. 4-11). These features can best be understood by considering these simplifications:

1. In the middle and coastal domains mean advection is negligible.
2. Water mass transformations occur locally, primarily through heat and salt transfer at the surface.
3. Vertical profiles are determined by the interplay of buoyancy addition and mechanical stirring, and in the outer domain also by lateral interleaving between shelf and oceanic waters.
4. Rates of buoyancy addition change annually, while stirring (primarily tidal) remains nearly constant with time and increases shoreward.

During winter, the separation into these domains is less clear, and the addition of negative buoyancy and stronger wind stirring move the boundary of vertical homogeneity seaward through the middle shelf. Even during this season, however, the potential for stratification like that in summer remains, and melting ice can provide sufficient buoyancy to stratify waters in the middle domain.

The hydrographic structure influences mixing, and the system of domains and fronts affects many distributions: e.g., salt, heat, momentum, vorticity, sediment, benthos, plankton, nutrients, fish, and

pollutants. With few exceptions, we do not understand these effects, and in many cases we do not even know what the effects are. As we have suggested, some effects of salinity and temperature distributions and their interactions with the hydrographic structure are straightforward, but many others are not. Future studies of the shelf will have to consider the influence of hydrographic structure on many phenomena.

Figure 4-11. A schematic of the cross-shelf density structure illustrating the system of hydrographic domains separated by fronts. This picture represents summer conditions, when the structure is clearest. Vertical profiles are shown beneath each domain. See Tables 4-2 and 4-3 for a tabulation of domain properties.

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Circulation Over the Continental Shelf of the Southeastern Bering Sea

5

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ABSTRACT

Using extensive direct current measurements made during the period 1975-78, we describe flow over the southeastern Bering Sea shelf. Characteristics of the flow permit us to define three distinct regimes, nearly coincident with the hydrographic domains defined in the previous chapter. The coastal regime, inshore of the 50 m isobath, had a slow (1-5 cm/sec) counterclockwise mean current and occasional wind-driven pulses of a few days' duration. The middle regime, bounded by the 50 and 100 m isobaths, had insignificant (<1 cm/sec) mean flow but relatively stronger wind-driven pulses. The outer regime, between the 100 m isobath and shelf break (~170 m), had a 1-5 cm/sec westward mean and low-frequency events unrelated to local winds. Over the entire shelf most of the horizontal kinetic energy was tidal, varying from 60 percent in the outer regime to 90 percent in the coastal regime. About 80 percent of the tidal energy was semidiurnal.

Mean flow over the shelf is well described qualitatively by dynamic topographies, and shallow current data from both coastal and outer regimes agree quantitatively as well. Two meteorological conditions that force the observed current pulses have been identified. In summer eastward-traveling low atmospheric pressure centers caused low-frequency pulses in the middle regime, and weaker pulses in the coastal regime. In winter, outbreaks of cold and dry continental air forced pulses within the coastal and middle regimes.

INTRODUCTION

Until recently, few direct current measurements were available over this shelf, so that ideas about flow were based partly on indirect methods and partly on intuition (see the historical review which follows). Since 1975, however, we and many colleagues have gathered numerous direct current measurements with

concurrent hydrographic data. We are thus able to base our characterization of the flow over this shelf on plentiful information. Our analysis of this suite of data (Table 5-1) is still incomplete. We expect further analysis building on this preliminary report to improve understanding, but not to change fundamentally the conclusions that we present here.

The most important discovery as a result of these new data is the existence of three distinguishable flow regimes over the shelf. These flow regimes correspond closely to the hydrographic domains outlined in the preceding chapter. The coastal regime is shoreward of the 50 m isobath, the middle regime is between the 50 m and 100 m isobaths, and the outer regime is between the 100 m isobath and the shelf break. Seaward of the shelf break lies the oceanic regime. Although they nearly coincide, for clarity we refer to flow *regimes* and hydrographic *domains* in this and the previous chapter.

We emphasize the characteristics of these regimes as we examine the frequency distribution of the horizontal kinetic energy, the mean circulation, and seasonal variations. Before discussing our findings, we review the physical setting, highlight earlier work, and discuss our measurements.

Setting

The southeastern shelf is bounded by the Alaska Peninsula, the Alaskan mainland, the shelf break

running from Unimak Pass to the Pribilof Islands, and an arbitrary line running from the Pribilofs to Nunivak Island (Fig. 5-1). The shelf break occurs at an average depth of 170 m (Scholl et al. 1968), and the shelf shoals gradually over a featureless expanse of 500 km. Flow over the continental slope is highly variable (Kinder et al. 1975, Coachman and Charnell 1979, Kinder et al. 1980) with mean flow to the northwest at 5-10 cm/sec. The shelf regime is separated from the adjacent oceanic regime by a weak haline front (Kinder and Coachman 1978, Coachman

and Charnell 1979). We have found no convincing evidence for exchange of mass or momentum between the shelf and oceanic regimes by eddies or rings. This characteristic is probably the effect of some combination of the width of the shelf, the front overlying the slope, and the weakness of the boundary current above the slope. At the same time, a considerable volume of water must flow across this long (~1,000 km) shelf break somewhere: about $1 \times 10^6 \text{ m}^3/\text{sec}$ flows northward through the Bering Strait into the Chukchi Sea (Coachman and Aagaard,

Figure 5-1. Locations of current-meter moorings. Most moorings had two current meters, one 10 m above bottom and one 20 m below the surface. BC signifies Bristol Bay project and FX signifies Frontal Experiment. Sequential deployments at a mooring site are designated by letter suffixes, e.g., BC-9A. The three moorings southwest of Nunivak Island numbered 1-3 have FX prefixes; all others have BC prefixes.

Chapter 8, this volume) and river runoff and precipitation account for only about 2 percent of this. There is strong evidence that little of this required flow occurs southeast of the Pribilofs, and hydrographic distributions suggest that it occurs near Cape Navarin on the Siberian coast.

As we have explained more fully in the previous chapter, the shelf has three distinct hydrographic domains delimited by the 50 m and 100 m isobaths and by the shelf break (170 m isobath). Freshwater runoff from rivers at the coast and an excess of precipitation over evaporation maintain a surface salinity difference across the shelf of about 20/oo. First-year ice covers most of the shelf during winter, and the weather also changes with the season: these changes are discussed more fully in the chapters on ice (Pease, McNutt, this volume) and on weather (Overland, Niebauer, and Ingraham, this volume).

In summary, five factors influence shelf circulation:

- (1) The width of the shelf and the weak haline front over the slope separate much of the shelf from the adjacent oceanic water masses.
- (2) There is no strong boundary current adjacent to the shelf.
- (3) The shelf has three distinct hydrographic domains.
- (4) River runoff maintains a salinity gradient across the shelf.
- (5) Weather and ice cover vary annually.

HISTORICAL REVIEW

Until recently, investigators faced the serious problem that there were few data from which inferences about circulation could be made. And yet the need existed for those conducting such non-physical studies as geological, biological, and fisheries studies to know the circulation. For this reason circulation schemes were proposed based on various mixtures of current measurements, hydrographic measurements, drifter measurements, and large doses of intuition. It is not surprising that many of these inferences do not compare favorably with our present knowledge, but some ideas resulting from these early attempts do agree with our present ideas. There are long reference lists of earlier work in Arsenev (1967), Ohtani (1973), and Favorite et al. (1976).

Before 1975 direct current measurements virtually did not exist. A single 24-hour current measurement taken close to Nunivak Island during August 1934 indicated northward flow at 17 cm/sec (Barnes and Thompson 1938). During the summers of 1955 and

1956 the *Tokei Maru* attempted current measurements north of the Alaska Peninsula, but mean currents were obscured by the tides (International North Pacific Fisheries Commission 1957). Hebard (1961) reported June 1957 current measurements at four sites. Each measurement lasted 39 hours, and he inferred a cyclonic (counterclockwise) mean flow around Bristol Bay. Kinder and Coachman (1975) placed three current-meter moorings in Pribilof Canyon for two weeks during July 1974. They measured vector mean speeds from 0.6 to 10.8 cm/sec, and directions mostly parallel to the local isobaths. Some efforts were also made to infer currents from drift bottle experiments (e.g., Thompson and Van Cleve 1936, Dodimead et al. 1963), but little was learned from the few recoveries.

Attempts were also made to describe the circulation using dynamic topographies. On the face of it these attempts seem forlorn, for neglected forces (e.g., friction, sea surface slope, wind stress) may be more important than those retained (i.e., Coriolis and the pressure gradient caused by varying density). In spite of this, mean flows on shelves are often qualitatively described by dynamic topographies, and some features in the Bering topographies agree with direct measurements. Natarov (1963) showed a surface topography (relative to 1,000 db¹) with broad northward flow across the shelf for summer 1959. Arsenev (1967) constructed a similar map of summer surface current based on 1,000 db. His map showed about 10 cm/sec cyclonic circulation in Bristol Bay, two eddies farther seaward, and northeastward (i.e., shoreward) flow between the Pribilofs and Nunivak Island. Adding wind-driven (Ekman) current weakened one eddy and changed the Pribilofs-Nunivak flow to northwestward. Neither Natarov nor Arsenev explained their method for extrapolating dynamic topography relative to 1,000 db from the deep basin across the shelf. Ohtani (1973) also calculated a surface dynamic topography, using a reference level of 50 db (his calculations did not extend inshore of the 50 m isobath). He showed a broad northwestward flow, but cautioned that such topography has limited application in shallow water.

Attempts were also made to deduce the flow from examination of hydrographic properties. Kitano (1970) apparently thought that the cold bottom water over the middle shelf in summer was advected southeastward from the Gulf of Anadyr (northwestern Bering Sea shelf). Conversely, Takenouti and Ohtani (1974) showed flow in the opposite direction

¹ One decibar (db) is approximately equivalent to one meter depth.

through the middle shelf based on their perception of shelf water masses.

Finally there were circulation schemes, opinions of various authors based on the meager useful data available and their own intuition. The U.S. Navy Hydrographic Office (1958, reproduced by Hughes et al. 1974 as their Fig. 3.4) reported mean currents of 0.1- 0.5 knots (5-25 cm/sec) across the southeastern shelf based on ship drift; a later publication (Naval Oceanographic Office 1977) revealed that this erroneous result was not based on data (Brower et al. 1977 showed a similar circulation in their Figs. 3 and 4). The most ambitious attempt was by Favorite et al. (1976, their Figs. 41 and 42). They hypothesized mean flow paralleling the shelf break (transverse current, identical to Kinder, Coachman, and Galt's 1975 Bering Slope Current), flow onto the shelf paralleling the Alaska Peninsula (West Alaska Current), cyclonic flow around Bristol Bay, no flow over most of the middle shelf, and a southwestward current from near Kuskokwim Bay toward the Pribilofs (Pribilof Current). They did not assign speeds to these flows, but recent data support their general scheme, except for the Pribilof Current.

In general, early attempts to describe flow over the shelf were unsuccessful and even deceptive. The few direct measurements used were overwhelmed by tidal currents and of too short duration to resolve the mean. Dynamic heights referred to 1,000 db were alleged to show flow over the shelf. Local transformations of water masses were ignored and distributions of temperature and salinity were attributed solely to advection. Circulation schemes were drawn showing more detail than the evidence justified. We present our own circulation scheme of the mean flow later in this chapter but we hope that it is used with two questions in mind: What is the *statistical* significance of this depiction of the mean flow? What is the *physical* significance of the mean flow?

METHODS

Measurements

Our inferences are based on 60 current-meter time-series records made during 1975-78 (Table 5-1, Fig. 5-1). Usable record lengths varied from 9 to 246 days, and a total of about 16 record-years of data were obtained. All data were acquired by RCM-4 Aanderaa current meters on taut-wire moorings. Typical instrument locations were 20 m below the surface and 10 m above bottom, as we avoided both the most active part of the surface layer and the bottom boundary layer. Since we also wanted the shallowest

instrument on each mooring to be above the seasonal pycnocline, which averaged about 20 m deep, shallow instrument placement was a compromise. Much has been written about the peculiarities of these instruments, and their possible errors due to "rotor pumping," mooring motion, and rotor stalling (Quadfasel and Schott 1979 give several references). In our measurements these errors were minimized by:

- (1) use of subsurface flotation (although the upper float was typically within 20 m of the surface);
- (2) short mooring length, all less than 200 m and typically less than 100 m;
- (3) use of taut moorings, about 1,000 lb (1 lb = 4.45 newton) buoyancy;
- (4) large and persistent (tidal) scalar speeds, preventing rotor stalling;
- (5) large tidal currents, which maintained relatively steady vane alignments.

All the data that we use have been averaged in some manner, so that random errors are not a serious problem. Biases will not average out, and we estimate direction accurate to $\pm 5^\circ$ and speeds within ± 1 cm/sec (exclusive of rotor pumping and mooring motion). Mayer et al. (1979) discussed similar moorings, and found an erroneous increase in energy up to two-fold (i.e., 40 percent speed increase) in winter on the middle Atlantic shelf. In the Bering Sea strong summer storms are infrequent, and ice cover severely damps surface waves during much of the winter, so that these estimates seem too high for our moorings.

We estimated the speed error by examining the most energetic tidal constituent, the principal lunar semidiurnal (M_2). From records BC-9B and BC-20B (see Table 5-1 and Fig. 5-1) we took one-month segments from the shallow (23 m and 22 m depth) instruments. We compared periods when the mooring was beneath ice cover to periods when the surface was free of ice, assuming that the ice damps surface waves (e.g., Wadhams 1978) so that there was no error induced by waves during the period of ice cover. We further assumed that the M_2 constituent was unchanged by ice cover (it may be affected, but it is probably diminished by ice, making our calculations conservative; see Chapter 8, Pearson et al., this volume). We found a 19-26 percent speed increase during the ice-free season. Although the errors probably vary significantly with weather, we believe that our speeds are accurate to about 25 percent near 20 m depth, and the speeds from deeper instruments are probably more accurate than this.

We also used drifters, floating buoys with "window shade" drogues at 10 m or 17 m depths. The tracks of these surface buoys closely correspond to the tra-

TABLE 5-1

Current-meter records

Mooring	Water depth	Meter depth	Scalar speed	Vector mean speed	Direction	Record length	Period
	(m)	(m)	(cm/sec)	(cm/sec)	(° T)	Days	
COASTAL REGIME							
FX-1A	48	38	21.2	1.7	294	63	7/20-9/20/78
FX-2A	43	20	32.4	1.9	306	63	7/19-9/20/78
FX-3A	46	14	33.1	2.3	319	63	7/20-9/20/78
		36	21.9	1.7	295	63	7/20-9/20/78
BC-9A	41	17	28.0	0.1	248	60	6/2-7/31/76
		27	24.8	0.4	251	60	6/2-7/31/76
BC-9B	41	23	28.6	4.4	309	231	9/24/76-5/12/77
		33	16.1	2.9	311	231	9/24/76-5/12/77
BC-9C	41	23.5	24.6	1.0	316	122	5/12-9/10/77
		33.5	19.3	0.9	279	120	5/12-9/8/77
BC-14A	51	20	34.9	1.4	086	31	5/29-6/28/76
		20	33.1	4.8	065	33	8/27-9/28/76
		37	27.4	2.1	211	31	5/29-6/28/76
		37	20.6	0.6	345	33	8/27-9/28/76
BC-15A	50	20	28.8	2.4	274	118	5/31-9/26/76
BC-15C	50	20	29.9	2.2	273	130	5/4-9/10/77
		34	22.4	1.0	304	130	5/4-9/10/77
BC-16A	49	20	28.2	0.5	314	131	5/3-9/10/77
		37	21.2	0.2	011	131	5/3-9/10/77
BC-18A	31	20.5	24.5	1.2	011	122	5/12-9/12/77
BC-19A	28	22.5	27.6	0.8	156	12	5/12-5/23/77
MIDDLE REGIME							
BC-2A	65	20	17.2	0.5	306	58	9/8-11/5/75
		50	10.0	0.9	301	59	9/8-11/5/75
BC-2B	65	50	17.6	1.2	089	192	11/5/75-5/14/76
BC-2C	65	20	14.8	0.8	324	119	5/31-9/26/76
BC-2D	65	21	17.1	1.9	080	194	9/27/76-4/8/77
BC-2E	65	20	16.9	0.7	214	130	5/4-9/10/77
		50	15.3	0.3	251	130	5/4-9/10/77
BC-4A	55	30	29.3	2.7	299	58	9/8-11/4/75
		47	21.9	0.6	311	58	9/8-11/4/75
BC-4B	55	30	25.4	2.0	301	209	11/5/75-5/31/76
BC-4C	55	25	27.8	2.7	312	61	6/1/76-7/31/76
		52	19.5	1.3	299	61	6/1/76-7/31/76
BC-4D	55	20	31.9	3.6	310	119	9/25/76-1/21/77
		48	18.2	3.0	313	174	9/25/76-3/17/77
BC-4E	55	20	27.7	1.1	300	64	5/13/77-7/15/77
		48	18.3	0.4	274	45	5/13/77-6/28/77
BC-4G	55	18	30.1	3.9	294	64	7/19-9/20/78
		46	16.9	2.1	302	64	7/19-9/20/78
BC-5A	70	20	19.6	1.2	253	31	5/29-6/28/76
		20	15.5	0.2	350	33	8/27-9/28/76
		50	27.8	0.9	298	31	5/29-6/28/76
		50	31.6	1.8	170	33	8/27-9/28/76
BC-6A	76	20	11.3	0.6	012	123	5/29-9/28/76
		50	23.3	0.7	023	31	5/29-6/28/76
		50	18.2	1.4	059	33	8/27-9/28/76
BC-8A	73	26	21.8	0.9	089	60	6/1-7/31/76
		54	21.9	0.6	348	60	6/1-7/31/76

TABLE 5-1, cont.

Mooring	Water depth	Meter depth	Scalar speed	Vector mean speed	Direction	Record length	Period
BC-10A	66	49	21.9	2.0	171	68	6/1-8/8/76
BC-12A	95	39	9.2	0.6	278	84	3/19-6/11/76
OUTER REGIME							
BC-3A	115	20	29.0	3.2	011	130	11/7/75-3/16/76
BC-3B	116	25	27.8	2.8	334	9	3/17-3/25/76
		105	17.4	1.2	342	73	3/17-5/28/76
BC-3C	114	20	31.8	17.2	012	123	5/29-9/28/76
		100	20.5	6.7	000	123	5/29-9/28/76
BC-13A	122	20	20.6	3.3	353	69	3/22-5/29/76
		100	11.9	1.6	348	87	3/22-6/16/76
BC-13B	115	100	17.3	1.6	335	36	6/6/76-7/12/76
BC-13C	108	22	25.2	5.3	333	202	9/29/76-4/19/77
		96	16.1	4.4	321	83	9/29-12/21/76
BC-17A	104	96	18.1	3.2	296	142	9/22-3/11/77
ST. MATTHEW							
BC-20A	64	22	23.8	3.4	335	157	9/17/77-2/20/78
		52	16.7	1.4	356	177	9/17/77-3/13/78
BC-20B	64	51	14.6	2.8	334	54	7/21-9/13/78
BC-21A	42	29	32.4	1.5	292	246	9/16/77-5/20/78
BC-21B	42	20	28.8	2.3	338	55	7/20-9/13/78
		32	20.3	0.8	317	55	7/20-9/13/78

The prefix (i.e., BC, FX) indicates the project (Bristol Bay or Frontal Experiment), and indicates the number of the mooring location. The suffix refers to a particular deployment, and individual instruments are identified by their depths. Thus BC-9B 20 m refers to a record obtained at 20 m depth from the second deployment at mooring 9 by the Bristol Bay Project (Fig. 5-1 shows locations). Some records are separated because of midsummer biological fouling problems encountered near the Alaska Peninsula (e.g., BC-14A).

jectories of water parcels at drogue depths. Using the Nimbus satellite, the position of these drifters was fixed about four times daily. This time series of positions (error = ± 4 km) was then processed to yield velocities. Kinder et al. (1980) discuss this technique in more detail.

Processing

Sample intervals of the current meters varied from ten minutes to one hour, depending on the expected length of deployment. All data were filtered, with one or two low-pass filters (i.e., they pass low frequencies and block higher ones). All data were originally filtered with a three-hour filter to minimize high-frequency noise. To examine lower frequencies, a further 35-hour low-pass filter was used to remove the tides. Table 5-2 lists basic filter properties, and Charnell and Krancus (1976) discuss processing details.

The 3-hour low-pass filter removes unwanted

(for our purposes) high-frequency data, while the 35-hour filter removes diurnal and semidiurnal signals while leaving periods longer than about two days intact. Note that 50 percent amplitude corresponds to 25 percent kinetic energy; the commonly used "half-power point" occurs at longer periods (lower frequencies) than the 50 percent amplitude point.

TABLE 5-2

	Filter properties		
	Period (hours) with given % amplitude remaining		
	0.1%	50%	99%
3-Hour Filter	2	2.9	5 hours
35-Hour Filter	25	35	55 hours

FREQUENCY DISTRIBUTION OF HORIZONTAL KINETIC ENERGY

One of the most useful ways of examining time-series data of water velocity is to calculate the kinetic energy distribution as a function of frequency. Some of the ocean's processes occur at distinct frequencies, or in discernible frequency bands: often a chaotic jumble in plots of velocity versus time becomes clear when energy versus frequency is plotted. The most striking example is the tidal currents, whose frequencies are determined by astronomical constants.

For the southeastern shelf, we define three frequency categories: mean, low-frequency (subtidal), and tidal. The mean flow is the vector average over a few months or longer. We have in mind the flow over a season or longer, but in practice we have usually defined the mean by the length of the current record. Two tidal periods stood out, diurnal (about 24-hour period) and semidiurnal (about 12-hour period). Low-frequency (long period) flow then fell between seasonal and daily periods. Although there were no definite periods associated with the low-frequency energy, one week was typical. In most records, these three frequency categories contained over 90 percent of the kinetic energy. Tidal frequencies contained most of the energy over the shelf, ranging from 60 percent of the fluctuating energy² near the shelf break to more than 95 percent in some records obtained inshore of the 50 m isobath. Roughly 80 percent of this tidal energy was semidiurnal, and 20 percent diurnal (Pearson et al., Chapter 8, this volume, describe the tides more fully). Vector mean speeds were usually 10 percent or less of the mean scalar speeds (Table 5-1), so that the kinetic energy of the mean flow was about 1 percent of the total horizontal kinetic energy ($KE = MV^2/2$, or per unit mass $KE/M = V^2/2$). Of the 60 records in Table 5-1, only three had vector mean speeds exceeding 5 cm/sec, while only six records had scalar mean speeds below 15 cm/sec. Low-frequency flow, that appearing at frequencies between those of tidal and mean flow, accounted for 3-20 percent of the energy. Over the outer shelf, energies in this frequency band were higher than farther inshore. Frequencies in these bands matched frequencies of weather phenomena, say 2-10 days (see weather chapters), and of the inherent variability of the mean flow over the outer shelf and slope (e.g., eddies and meanders). The

² The fluctuating kinetic energy (per unit mass) is the total kinetic energy, less the kinetic energy of the mean ($FE = \frac{1}{2}(U - \bar{U})^2$). It is often referred to simply as the energy or total energy and is equal to one-half the variance.

inertial period (about 14 hours at 58° latitude), which is important in many open-ocean records of kinetic energy, accounted for only 1 percent or less of the total energy in our records. Table 5-3 illustrates the frequency distribution of the horizontal kinetic energy in some typical records.

Basic plots

We can illustrate the character of flow over the shelf most clearly by showing plots of typical velocity records. Although there were significant differences between records and between regimes, we attempt to highlight common features of the records.

Fig. 5-2 shows unfiltered data from BC-15A (Fig. 5-1) at a depth of 20 m during September 1976.³ Samples are plotted every 20 minutes (the current-meter sample interval) as two components. The upper plot is the east-west component (U), with eastward flow positive, and the lower plot is the north-south component (V), with northward flow positive. Two features are obvious from this depiction. First, a tidal signal overwhelmed any other variable signals present, and it was mostly semidiurnal (about 12-hour period). This record came from the coastal regime, where the tides contributed 90 percent of the fluctuating kinetic energy, and where the semidiurnal tide typically accounted for 80 percent of the tidal energy. Second, the upper plot is displaced below the zero axis, while the lower plot is centered about this line. Thus, for the week shown,

Figure 5-2. A portion of BC-15A 20 m in September 1976. This record is unfiltered, and the sampling interval was 20 minutes. A. East-west (U) velocity component and B. North-south (V) velocity component. Strong semidiurnal and less strong diurnal tides show clearly. The U component is displaced below the axis, indicating a mean flow to the west.

³ The draftsman has inadvertently low-pass filtered this record somewhat. Original plots made with fine pens show considerably more jiggles (high-frequency energy).

TABLE 5-3
Frequency distribution of horizontal kinetic energy (cm^2/sec^2)¹

Record	Mean	Fluctuating ²	Low	Diurnal	Semi-diurnal	Inertial
Coastal Regime						
BC-9B, 23 m, winter	2(0)	434(100)	59(14)	74(17)	271(62)	1(0)
BC-9C, 23 m, summer	1(0)	340(100)	6(2)	73(21)	256(75)	0(0)
Middle Regime						
BC-2B, 50 m, winter	1(0)	203(100)	11(5)	68(33)	120(59)	0(0)
BC-2C, 20 m, summer	0(0)	175(100)	3(2)	49(28)	110(63)	1(1)
Outer Regime						
BC-3C, 100 m, summer	20(8)	266(100)	30(11)	42(16)	165(62)	2(1)
BC-3C, 20 m, summer	148(29)	510(100)	149(29)	60(12)	251(49)	7(1)
BC-3A, 20 m, winter	5(1)	511(100)	112(22)	65(13)	285(56)	6(1)

¹ cm^2/sec^2 is a unit of energy per unit mass, i.e., erg/g ($1 \text{ erg/g} = 10^{-4} \text{ J/kg}$).

² Fluctuating kinetic energy is the total kinetic energy in the records, less the kinetic energy of the mean. It is one-half the record variance.

Note that the semidiurnal tidal energy does not change much between summer and winter. Percent of fluctuating energy is shown in parentheses.

the mean current was westward (cf. record mean of 2.4 cm/sec at 274°T). The entire BC-15A record had one of the clearest and most consistent means of all those that we have examined, but usually attempting to measure average currents in flows like this with records of a few days would be futile.

Fig. 5-3 shows a different presentation, progressive vector diagrams (often abbreviated PVD), all smoothed with the three-hour low-pass filter. Progressive vector diagrams show what the trajectory of a water parcel *would* have been *if* it had had the velocity measured at the site of the current meter. This is much different from familiar depictions of flow such as trajectories (paths of particles) or streamlines (lines parallel to velocity vectors). Fig. 5-3 shows progressive vector diagrams from each of the three regions: outer (BC-13A 20 m), middle (BC-6A 50 m), and coastal (BC-15A 20 m). The record mean was largest at the outer mooring (note different distance scales), and least at the middle mooring. All records had vigorous clockwise tidal motion, which was nearly circular in the inner record, and elliptical in the middle record (major axis parallel

to the local isobaths). During much of the outer record the tidal circles were superposed on strong low-frequency flow so that they appear as cusps. These subtidal frequency pulses typically lasted for a few days (time may be inferred in Fig. 5-3 by counting clockwise semidiurnal circles), and can be seen in all three records. In the outer record, both speed and direction of pulses were variable. The middle and coastal records showed pulses that appeared similar at each occurrence, northward at BC-6A and westward at BC-15A. Our division of the flow into three frequency bands (mean, low-frequency, tidal) can be seen in each of these examples, and the examples also emphasize the unsteadiness of the flow.

Low-frequency plots

To examine the low-frequency (subtidal) flow, the 35-hour low-pass filter was applied to records before plotting. Plots of components versus time (Fig. 5-4A, 4B) and stick plots (Fig. 5-4C) were then made. Stick plots are a time series of vectors, the base of the

Figure 5-3. Progressive Vector Diagrams (A) Outer regime, BC-13A 20 m. (B) Middle regime, BC-6A 50 m. (C) Coastal regime, BC-15A, 20 m. These figures are the displacement of a water parcel having the same velocity as measured by the current meter, and may be quite different from the actual trajectory of any parcel. The outer record showed tides (cusps), strong pulses, and a northward mean. The middle record showed strong tides (clockwise ellipses), weak northward pulses, and perhaps a weak northeastward mean. The coastal record showed strong tides, westward pulses (extending the tidal circles like a loose spring or slinky toy), and a strong westward mean. Note different scales. (S signifies the start and F the finish of each plot.)

vector indicating the time, the direction away from the base indicating the direction toward flow, and the length of the vector indicating speed. The record from BC-15A at a depth of 20 m shows generally westward flow at 5 cm/sec, with higher speeds later in the record. We found that during the generally low wind speeds (<5 m/sec) in summer there was little correlation between wind and current. During strong winds (>10 m/sec), especially in autumn when higher winds are common and we have many records, wind and currents were much better correlated. The record segments with close correlation seemed to occur for two to four days during the passage of atmospheric low-pressure centers through the area.

Spectra

Except for truly dominant frequency components (e.g., tides), it is difficult to describe the frequency distribution of the records from the plots we have presented thus far. A useful technique for doing this (and the one used to make the estimates for Table 5-3) is spectral analysis. A time series is mathematically (Fourier) transformed, so that instead of velocity as a function of time, we have kinetic energy as a

Figure 5-4. Low-frequency flow. Low-pass (35-hour) filtered data from BC-15A 20 m. (A) East-west component, (B) North-south component, and (C) Vector sticks (cm/sec). Much of the low-frequency flow occurred during strong (>10 m/sec) wind events.

function of frequency. When plotted the spectrum gives a concise picture of the frequency distribution of the kinetic energy (variance). Of course, the characteristics that showed clearly on other plots should also be evident in the spectral plots.

Calculating the spectrum for one series (autospectrum) can be useful, but this can also be done for two records (cross spectrum). Normalizing such a spectrum by the variances produces a "coherence" as a function of frequency. The coherence is the frequency-dependent counterpart of the correlation function between two series as a function of time lag; instead of showing the time lags at which records are correlated, it shows the frequencies at which records are phase locked (the calculation also yields the corresponding phase spectrum, a frequency-dependent time difference). We plot the square of coherence because, like the squared correlation coefficient, the squared coherence corresponds to the percent of the variance at the frequency that is "explained" by a linear relationship. Coherence is always positive, but phase information (not shown) conveys sign. For instance, two sine waves of identical frequency but opposite sign have a coherence of 1.0 and a phase of 180° .

We illustrate with examples that summarize the frequency distribution of kinetic energy over the

shelf. Autospectra for BC-15C, 20 m and 34 m depth, from summer 1977 are plotted in Fig. 5-5. The vertical axis is kinetic energy density, i.e., per unit frequency, $(\text{cm/sec})^2/\text{CPD}$ (CPD is cycles per day). In this presentation the vector series have been decomposed into clockwise (anticyclonic) and counterclockwise (cyclonic) rotating components, instead of into north-south and east-west components. When there is no strong local orientation (e.g., a shoreline or strong bathymetric gradient), the clockwise/counterclockwise decomposition is more useful than the east/north decomposition. The most striking feature of both records is the high diurnal (1 CPD) and semidiurnal (2 CPD) peaks, with the semidiurnal higher. This conforms to our knowledge that motion over the shelf is tidally dominated (cf. Figs. 5-2 and 5-3). For the shallow record 89 percent of the energy was tidal (70 percent semidiurnal) and for the deeper record 94 percent was tidal (67 percent semidiurnal). At low frequencies there were small irregular peaks. This also corresponds to expectation, as the low-frequency filtered records did not reveal any dominant frequencies (cf. Fig. 5-4). Near the inertial frequency (1.7 CPD), in the clockwise components only, there is a small, statistically insignificant peak. One characteristic of inertial

Figure 5-5. Autospectra, BC-15C 20 m and 34 m. The energy density ($\text{cm}^2 \text{sec}^{-2} \text{CPD}^{-1}$) versus cycles per day shows strong diurnal and semidiurnal tidal peaks, and irregular low-frequency peaks. Clockwise (solid) and counterclockwise (dashed) current components are plotted.

oscillations (in the Northern Hemisphere) is clockwise rotation, and the presence of similar peaks near the inertial frequency in most records has led us to label these peaks inertial. An example of vertical coherence from mooring 5A at 20 m and 50 m depth during summer 1976 is shown in Fig. 5-6. Again, we used clockwise and counterclockwise components. Not surprisingly, the tidal peaks show clearly. Tidal motion at 20 m depth was coherent with motion at 50 m depth. In our records vertical coherence >0.9 at tidal frequencies was common, and since most of the energy was at tidal frequencies, this implies that shallow and deep motions were generally coherent. As with most of the records there are also some low-frequency peaks above the 95 percent confidence limit (~ 0.26). Examining individual records (plotted versus time), we often found that shallow and deep records were highly correlated during strong winds, but over the entire record low-frequency flow was often weak and uncorrelated. This is consistent with coherence estimates like that in Fig. 5-6.

In general then, we find that motion throughout the sampled water column was highly coherent at tidal frequencies, and either weakly coherent or not

Figure 5-6. Vertical coherence squared. The squared coherence of BC-5A 20 m with BC-5A 50 m is shown, in clockwise (solid) and counterclockwise (dashed) components. Highly significant coherences in tidal bands are common to all records. Some also show strong low-frequency coherence and inertial (1.7 CPD, clockwise) coherence.

coherent at other frequencies. When the coherence was high, the phase differences were usually small, so that the flow at these frequencies was in the same direction at both depths. During strong winds, the low-frequency flow was vertically correlated (see below, Seasonal Variations and Meteorological Forcing) and this probably caused the occasional peaks in coherence at low frequencies. Coherence calculations do not reveal relative speeds, but an examination of Table 5-1 shows that shallow instruments usually recorded faster speeds than deeper ones. This may be instrumental (see Introduction) or it may reflect an actual decrease with depth.

Just as it is possible to calculate the vertical coherence to help determine whether motions are similar throughout the water column, horizontal coherences calculated between records from different moorings help determine if motions are similar over broad areas. The low-frequency sticks in data from July 1976 suggested to us that the water motion at low frequencies might be coherent over part of the shelf (see Fig. 5-10); we calculated coherence between BC-5A (50 m), BC-2C (20 m), and BC-6A (50 m). The deep instrument on mooring BC-2C failed, but the current meter at a nominal depth of 20 m was below the pycnocline. Calculations between these instruments showed coherence at separations of 30, 42, and 72 km. Again, the calculated tidal coherences squared exceeded 0.9 (Fig. 5-7). At low frequencies there was also high coherence squared in a broad peak which decreased with separation: at 30 km it was ~ 0.7 , and at 72 km it was ~ 0.55 . These peaks suggested that the middle shelf was similarly affected by low-frequency motions, and probably most of these motions were generated by wind events similar to those discussed under Seasonal Variations and Meteorological Forcing. Values of coherence squared between the middle shelf and coastal records at comparable distances (~ 60 km) were lower (~ 0.4), but still significant.

MEAN CIRCULATION

Although the kinetic energy distributions were dominated by tidal currents, many of the records had significant means (e.g., Fig. 5-3A). We examine the statistical significance of the record means, and construct a circulation scheme using them. The mean flow is energetically two orders of magnitude smaller than the tidal flow over much of the shelf, however, and the mean circulation may not be the most important flow component in many situations.

A vivid example is the flow through the Aleutian

HORIZONTAL COHERENCE SQUARED

Figure 5-7. Horizontal coherence squared. Even at a separation of 72 km, low-frequency (~ 0.3 CPD) coherence squared is nearly 0.6, and at 30 km it exceeds 0.7. The horizontal coherence was higher over the middle shelf (shown here) than in the coastal or outer regimes. Clockwise components have solid lines.

Passes (such as Unimak Pass). Flows in many of these passes have high mean scalar speeds, exceeding 100 cm/sec (see U.S.D.C. 1964), but these are mostly caused by tidal currents and may average to a vector mean of zero. Properties (e.g., salt and drift cards) can be exchanged between the North Pacific and Bering Sea through such a pass; it is unlikely that the same volume of water flows back and forth on ebb and flood, and the vigorous stirring near coast and bottom prevents such a volume of water from remaining intact. Practically, it is difficult to measure the mean flow in such passes with present techniques. For many oceanographic questions, the mean flow is not as important as the exchange of properties between two oceanic regimes and the mixing by vigorous tidal stirring.⁴

Just as obviously, the mean circulation over the Bering Sea shelf, even if much smaller than the tidal currents, can importantly influence many processes. For instance, plankton advected by a 2 cm/sec vector mean flow would travel about 150 km over the summer. Such a small mean flow is difficult to measure in a highly variable and active (noisy) background. The 2 cm/sec mean flow may be important, but separating it from the tidal and other variable currents with statistical significance requires measurements of long duration. For example, we had six

different moorings at location BC-4 (Fig. 5-1), and for each mooring we computed a mean (Table 5-1) from records 45-209 days long, and instruments 18-52 m deep. These means accounted for less than 1 percent of the horizontal kinetic energy in this tidally dominated flow. All means fell in the northwestern quadrant, varying from 274 to 311°T with speeds from 0.4 to 3.6 cm/sec. The unweighted means for these records ($\pm$ one standard deviation) were: 2.1 ± 1.2 cm/sec and $301 \pm 11^\circ\text{T}$. Although we cannot compare this to the actual mean, the repeatability gives us confidence in our data. Some of the variability was not due to measurement error: seasonal differences (wind stress, ice cover, and stratification), vertical velocity shear (the means of shallow records averaged 1.2 cm/sec faster than the deeper ones), and interannual changes accounted for some variation.

A reasonable qualitative estimate of the significance of the mean flow can be inferred from progressive vector diagrams (Fig. 5-3). To quantitatively estimate the statistical significance of the means, we used low-pass (35-hour) filtered records. We define a time scale⁵ (τ) of the low-frequency flow and then estimate the number of independent samples of the mean by dividing into total record length (T). Thus an estimate of the error (at 67 percent confidence) is:

⁴ Mean flow is often assumed through Unimak Pass into the Bering Sea; we know no evidence for this, only for an exchange between the Bering and the North Pacific.

⁵ The time scale is estimated by computing the area under the autocorrelation function; it is thus twice the "integral time scale" (Kundu and Allen 1976).

$$E = \frac{\sigma}{(T/\tau)^{1/2}}$$

E: root mean square error estimate (cm/sec)

σ : standard deviation of record (after low-pass filtering; cm/sec)

T: record length (sec)

τ : time scale (sec), the length of an independent sample.

An example of these calculations is given in Table 5-4, illustrating the significance of means for summer 1976. For these records, the mean was usually significant in the coastal regime, where it was about three times the estimated error. Over the middle

TABLE 5-4

Means and standard errors

Record	U ± Eu (cm/sec)	V ± Ev (cm/sec)	S, E ¹
Coastal			
BC-15A (20 m)	-2.3 ± 0.3	0.1 ± 0.5	2.3, 0.6
BC-14A (20 m)	3.3 ± 0.4	1.2 ± 1.0	3.5, 1.1
(40 m)	-0.4 ± 0.4	-0.6 ± 0.4	0.7, 0.6
Middle			
BC-2C (20 m)	0.4 ± 0.4	0.6 ± 0.3	0.7, 0.5
BC-5A (20 m)	-0.1 ± 0.5	-0.5 ± 0.4	0.5, 0.7
(50 m)	-0.1 ± 0.3	0.1 ± 0.7	0.1, 0.8
BC-6A (20 m)	0.6 ± 0.6	0.2 ± 0.4	0.6, 0.7
(50 m)	0.6 ± 0.4	0.8 ± 0.4	1.0, 0.6
Outer			
BC-17A (96 m)	-2.9 ± 0.1	1.4 ± 0.4	3.2, 0.4
BC-13C (22 m)	-2.4 ± 1.9	4.7 ± 1.5	5.3, 2.4

U and V are the record means, and Eu and Ev are the estimated standard errors (see text). S and E are the vector mean speed and the vector sum of the errors.

$$^1 S = (\bar{U}^2 + \bar{V}^2)^{1/2} \quad E = (Eu^2 + Ev^2)^{1/2}$$

shelf, however, the mean was generally equal to or less than the standard error. In the outer regime, the mean was much greater than the error. These examples conform to a general pattern: in the outer regime, means were greater than errors, in the coastal regime means were somewhat greater than errors, and in the middle regime means were insignificant. This generalization must be modified for measurements taken just seaward of the inner front separating the coastal and middle hydrographic domains (e.g., BC-4, Fig. 5-1). Such measurements show significant means because of their proximity to the front. These measurements also illustrate that the boundaries of the flow regimes nearly coincide with those of the hydrographic domains, but not exactly.

Computed means for each mooring location (Fig. 5-8) show the general circulation pattern. Over the entire shelf, significant mean flow usually paralleled local isobaths. In the outer regime mean flow was parallel to the shelf break, flowing northwestward at 1-5 cm/sec. In the middle regime mean flow was insignificant, usually <1 cm/sec. In the coastal regime between Cape Newenham and Nunivak Island flow was westward at 1-3 cm/sec, while along the Alaska Peninsula flow paralleled the nearby shoreline at 2-5 cm/sec.

Drifters, tracked by satellite, provided confirmation of flow character in the outer and middle regimes. Three drifters launched in June 1976 (near BC-2, -5, and -6) confirmed low vector mean speeds there, <1 cm/sec in the absence of strong winds. In 1977, six drifters released over the continental slope in the outer and oceanic regimes all drifted northwestward with vector mean speeds ~5 cm/sec (Coachman and Charnell 1979, Kinder et al. 1980). Two of these drifters wandered into the middle regime near the Pribilofs, where their vector mean speed was ~1 cm/sec. These drifters gave no evidence of exchange across the shelf break (i.e., crossing the front overlying the continental slope, Kinder and Coachman 1978), except in the area just north of the Alaska Peninsula.

Kinder et al. (1978) found water lying along the Alaska Peninsula below 20 m that was both warmer and more saline than surrounding waters. Muench and Ahlnäs (1976) noted that the region just north of the Alaska Peninsula was often free of ice when nearby water was ice-covered. These distributions could result from the advection of warmer oceanic water onto the shelf north of the peninsula. Moreover, one of the drifters deployed over the slope in 1977 crossed the shelf break north of the Unimak Pass, and traveled about 80 km eastward before grounding on the peninsula (Kinder et al. 1980).

Figure 5-8. Mean flow. The mean for all records at each mooring site is shown. Coastal and outer regime moorings generally had statistically significant means, while middle shelf sites did not. The domains refer to hydrographic structure (preceding chapter), but the domains and flow regimes are nearly coincident.

These measurements are strong evidence for flow across the shelf break near the Alaska Peninsula (note BC-14, Fig. 5-8), but hydrographic and current measurements also demonstrated that this flow is probably intermittent.

Mean flow over both the outer and coastal regimes appeared to be driven by pressure gradients caused by density differences. The method of inferring flow from dynamic calculations, referred to the deepest common depth of nearby hydrographic stations, gave approximate agreement with direct measurements (Coachman and Charnell 1979, Schumacher et al. 1979). Dynamic topographies in the outer regime (Kinder 1977, Coachman and Charnell 1979, Kinder

et al. 1978) indicated westward flow of ~5 cm/sec. Calculations across the inner front (near the 50 m isobath) yielded 1-2 cm/sec flow counterclockwise along the front and westward between Cape Newenham and Nunivak Island (Schumacher et al. 1979). Thus mean flow in both the outer and coastal regimes can be inferred from the density field, but this is not true of the middle shelf, where the mean is nearly zero. Dynamic topographies do suggest a very weak southeastward flow across the middle regime (e.g., Kinder 1977, Reed 1978, and Kinder et al. 1978), but the current records (Fig. 5-8) did not confirm this flow.

Synthesizing our knowledge of the shelf circu-

Figure 5-9. Estimated mean circulation. No strong distinction is made as to season or depth, although it is biased toward summer and the surface. The dashed arrows in the northern coastal regime express uncertainty, while the arrow along the Alaska Peninsula expresses intermittency. Flow over the shelf is mostly tidal, so that the instantaneous flow is quite different from this depiction. For instance, over the middle shelf we expect the speed at any time to exceed 20 cm/sec even though the long-term vector mean is less than 1 cm/sec, and instantaneous directions seldom agree with the arrows at any location. The scheme is also incomplete: the source of water for the westward transport in the coastal regime is not shown. We speculate about this source in the text.

lation, we have constructed a circulation scheme (Fig. 5-9). We emphasize again that the mean circulation is only a few percent of the total kinetic energy, and that even our large data sets do not define the mean flow adequately at some locations. That is, the arrows in Fig. 5-9 will often be incorrect instantaneously and scalar speeds will be much higher than the long-term vector means. Directions apply throughout the water column and speeds at the surface; speeds decrease significantly with depth seaward of the

100 m isobath. Our data are most extensive in summer (Table 5-1), but most data from other seasons agreed with this depiction.

In the coastal regime we show counterclockwise flow extending northward to St. Matthew Island with speeds of a few cm/sec. Within Kvichak and Bristol bays, measurements show that the counterclockwise flow extends to the coast (Straty 1977). The dashed arrow paralleling the Alaska Peninsula signifies intermittent flow. In the middle regime we show

weak mean flow (<1 cm/sec), direction indeterminate. In the outer regime and over the slope we show strong (1-10 cm/sec) flow paralleling the shelf break and flowing westward. Wherever the flow was well defined, the entire water column moved in the same direction, but speed decreased significantly with depth in the deeper water seaward of the shelf break. This flow is a continuation of eastward flow north of the Aleutian Islands which then curves cyclonically toward the west in the southeastern corner of the deep Bering Sea (Kinder et al. 1980). We also note that eddies seem to be common over the deeper water, so that "mean" currents there may radically change direction for weeks or months (Kinder and Coachman 1977, Kinder et al. 1980). We have not observed comparable eddies in the shelf waters.

We found no evidence of large flow onto the shelf; the flow leaving the shelf to the west in the coastal regime should also have small volumetric transport. We can estimate this outflow by assuming a 2 cm/sec flow over an average depth of 25 m along a section 150 km long, approximating conditions southwest of Nunivak Island to the inner front (i.e., spanning the coastal regime). This transport is $2 \text{ cm/sec} \times 25 \text{ m} \times 150 \text{ km} = 0.08 \times 10^6 \text{ m}^3/\text{sec}$, or less than 10 percent of the Bering Strait outflow ($\sim 1 \times 10^6 \text{ m}^3/\text{sec}$), but still larger than the river runoff ($0.0015 \times 10^6 \text{ m}^3/\text{sec}$; Roden 1967). Water to supply this westward flow may come from the middle shelf; across the middle regime, between the front and the Pribilofs, a southeastward transport of $0.08 \times 10^6 \text{ m}^3/\text{sec}$ across 200 km averaging 70 m deep would be only 0.5 cm/sec, detectable in neither our current nor our hydrographic data. Inflow along the coastal regime north of the Alaska Peninsula (30 km wide by 25 m deep) would need to be a steady 10 cm/sec, which is not supported by data, although some lesser inflow does occur there.

Thus the circulation scheme shown in Fig. 5-9 is incomplete because the source of water necessary to supply the westward transport in the coastal regime is not shown. We conjecture that this supply is mostly from the shelf northwest of our study region and that it flows southeastward into the middle regime. It may then enter the coastal regime in Bristol Bay, where the inner front appears weakest (see Chapter 4). Since the evidence for this circulation is weak, we exclude it from Fig. 5-9.

SEASONAL VARIATIONS AND METEOROLOGICAL FORCING

Some characteristics of the shelf environment change annually: insolation, river runoff, ice cover,

wind stress, and atmospheric pressure. Even though the tides, which have little seasonal variation (Pearson et al., Chapter 8, this volume), force most of the current, seasonal variation occurs in the non-tidal currents. These seasonal changes are small but still significant.

We showed some of the hydrographic changes in the preceding chapter. An example of changes in atmospheric forcing is shown in Table 5-5 (see also Overland and Niebauer, Chapters 2 and 3, this volume). During the winter, mean wind speeds are higher and storms are more frequent. Stratification is much weaker over most of the shelf, and we might thus expect more direct response to the wind in the absence of ice cover, which may alter the effect of the wind stress. Records from BC-2 and BC-9 indicate some seasonal changes, even when allowance is made for instrumental problems from more energetic surface waves (Table 5-6). Seasonal changes that we describe exceed the 25-30 percent error estimates (see Introduction).⁶

At BC-9 the shallow summer mean was 1.0 cm/sec toward 316°T . Decreasing the winter mean by one-third (a high estimate of the false inflation of speed by surface waves), the mean was 3.2 cm/sec toward 315°T . Likewise, BC-4 had higher vector mean speeds during winter, with the same direction as in summer. Another site with sufficient data to suggest

TABLE 5-5

St. Paul weather 1975-76

	Mean pressure (mbar)	Mean speed (m/sec)	Variance (m^2/sec^2)	Vector mean (m/sec, $^\circ$)
Summer 1976	1011	6.2	43	1.3 at 004
Winter 1975-76	1006	8.5	85	2.4 at 202

The weather station on St. Paul Island (northernmost of the Pribilofs) probably represents weather over the southeastern shelf better than the other stations, which are strongly influenced by topography.

⁶ The seasonal differences are slightly different from what might be inferred from Tables 5-1 and 5-3 because: Tables 5-1 and 5-3 use entire record lengths instead of smaller seasonal segments and we examine the meteorological frequency band (1.5-10 day) for seasonal effects, a subset of the low-frequency category in Table 5-3. A sequential 29-day tidal harmonic analysis was used to estimate errors in the Introduction, whereas Table 5-6 lists wide-band tidal estimates (identical with Table 5-3).

TABLE 5-6
Seasonal kinetic energy (cm^2/sec^2)

Record	Season	Fluctuating	Low ¹	Semidiurnal	Meteorological ¹
Coastal					
BC-9C 23 m	Summer 5/12-9/10/77	340	6	256	4
BC-9B 23 m	Winter 9/24/76-5/12/77	434	59	271	47
Middle					
BC-2C 20 m	Summer 5/31-9/26/76	175	3	110	2
BC-2B 50 m	Winter 11/5/75-5/14/76	203	11	120	7
Outer					
BC-3C 20 m	Summer 5/29-9/28/76	510	149	251	74
BC-3C 100 m	Summer 5/29-9/28/76	266	30	165	21
BC-3A 20 m	Winter 11/7/75-3/16/76	511	112	285	69

¹ Low-frequency energy includes all subtidal (less than diurnal frequency) energy in the record. Meteorological frequency energy is only that within the 1.5-10 day (0.67-0.10 CPD) band.

Note the relative constancy of the semidiurnal tidal energy at each mooring site.

seasonal changes is BC-2. Like BC-9 and BC-4, currents at BC-2 were faster in winter, with a vector mean speed (~ 1.5 cm/sec) about double the summer value. Unlike the other two sites, however, at BC-2 the direction reversed from westward ($\sim 300^\circ\text{T}$) in summer to eastward in winter.

The kinetic energy spectra showed that the greatest relative change occurred in the meteorological band, between 1.5 and 10 days (Table 5-6). This band corresponds to the increased variance seen in St. Paul wind data (Table 5-5). In the middle regime a record for 20 m in the summer of 1976 (BC-2) had meteorological kinetic energy of $2 \text{ cm}^2/\text{sec}^2$, while the deep 50 m record was threefold greater in the winter of 1976-77— $7 \text{ cm}^2/\text{sec}^2$. In the coastal regime 23 m records (BC-9) showed greater differences between

seasons. In summer 1977 meteorological band variance was $4 \text{ cm}^2/\text{sec}^2$, but during the winter of 1976-77 it was $47 \text{ cm}^2/\text{sec}^2$. These changes for the middle and coastal regimes exceeded our error estimates, but changes in the outer regime (BC-3) did not. This reinforces the idea that subtidal energy in the middle and coastal regimes is mostly meteorologically forced, while low-frequency energy in the outer domain is not forced by local weather. In the outer regime, the low-frequency energy is probably associated with variations in the Bering Slope Current.

Two different meteorological conditions can force low-frequency current pulses. One dominates the summer pulses and the second occurs only in winter. The summer condition is an eastward-moving low atmospheric pressure center, and the winter condition

is a southward outbreak of cold continental air. Strong storms associated with atmospheric low-pressure cells also occur in other seasons, but the continental air outbreak is exclusively a wintertime phenomenon.

Such a summer event occurred in late July 1976 when several current-meter moorings and drifters were placed in a line running northwest from the Alaska Peninsula (Fig. 5-10). (We suspect biological fouling of the upper instruments, especially BC-5 (20 m) and BC-6 (20 m), so that speeds in some records are too low.⁷) As the wind increased to 10 m/sec and rotated towards the east, currents at moorings 5, 6, and 2 responded similarly, with about

Figure 5-10. Current response to a summer wind event, middle and coastal regimes. Stick diagrams for current meters, drogued drifters tracked by satellite, and estimated wind. As the wind increased beginning on 27 July 1976, currents also increased and followed the wind in a clockwise rotation. Biological fouling probably decreased speeds at BC-5 20 m and BC-6 20 m.

⁷ In Chapter 28, F. Favorite cites an 1894 report of jellyfish fouling fishing gear near the Alaska Peninsula. Our fouling problem in the same area was also caused by jellyfish.

a one-day lag. That is, speeds increased, and direction rotated slightly clockwise toward the east after peak speeds were reached. The shallow current meter at mooring 14 showed a similar speed increase, but direction remained essentially parallel to the local bathymetric contours (Figs. 5-1 and 5-8). The two drifters still operating at that time had shown aimless wandering at mean speeds <1 cm/sec before the strong winds began. They were still near the moored instruments in late July (they were launched in early June), and responded strongly during this event. Both drifters accelerated to speeds near 20 cm/sec, and changed directions clockwise toward the east.

The current pulse illustrated by Fig. 5-10 was typical of many similar pulses in the current records, and is the best documented such event. These pulses also seem characteristic of the low-frequency currents in the two nearshore regimes. In our records, 50 percent of the low-frequency (subtidal) variance lies in the band 0.7-0.1 cycles per day (1.5-10-day period), which includes the repetition rate for storms.

The continental air outbreak, which occurs frequently in winter, apparently affects both the currents and ice distribution (see preceding chapter). High pressure over the Alaska mainland results in an offshore wind of cold, dry air blowing southward off the ice. This type of weather permits excellent satellite imagery (Fig. 5-11; see also Muench and Charnell 1977). Crossing the southern ice boundary, the cold, dry air receives heat and moisture from the underlying water. Clouds form downstream from the ice edge as long (hundreds of km) streamers paralleling the wind.

A record from BC-2B (Fig. 5-12) concurrent with a southward outbreak of cold continental air showed a strong (20 cm/sec) current pulse to the south during the period 19-22 January 1976. As the satellite image shows, winds were blowing SSE near BC-2 by 20 January in excess of 20 m/sec, and rotated slightly from south to south-southeast (cf. Fig. 5-10). Similar pulses were observed during winter 1977 at BC-2. In January, a five-day pulse averaged 15 cm/sec to the southeast and in February flow exceeded 30 cm/sec to the northeast for 30 hours.

Within the two inner regimes, winter brought energetic response to meteorological forcing (Table 5-6). Mean winds were stronger then (in the absence of storms), and strong winds occurred more frequently. Although the tides still dominated energy spectra, energy at meteorological frequencies played an increasing role, leading to the reversal of mean flow at BC-2. Ice cover probably modifies this forcing, but even under the ice (BC-9) wintertime winds cause current responses.

Figure 5-11. An infrared satellite image of the Bering Sea on 20 January 1976. The Alaska Peninsula, Pribilof Islands, Nunivak Island, and other land features show clearly. The ice pack extended south of Nunivak Island, with some typical lineations, mostly tending east-west and parallel to the ice front, near the edge of the ice. Thin streamer clouds, normal to the ice front and parallel to the wind, extended southward. Fig. 5-12 shows measured currents and estimated winds.

FLOW REGIMES AND HYDROGRAPHIC DOMAINS

We have used flow regimes as a framework for discussion in the preceding sections, and now we summarize their characteristics (Table 5-7). As with the nearly coincident hydrographic domains (preceding chapter), our description is biased toward summer. The discussion of seasonal variation showed, however, that many of the differences between regimes persist throughout the year.

Energy of the fluctuating flow was 90 percent tidal over the middle shelf, about 80 percent in the

coastal regime, but only 60-70 percent in the outer regime, where up to 30 percent of the energy was at periods greater than two days (Table 5-3). Inertial energy was about 1 percent of the energy in the outer regime, but barely detectable in the two shoreward regimes. The increase in tidal energy per unit mass in these two regimes is a consequence of shoaling. Larger fluctuating kinetic energy values at low frequencies in the outer regime probably reflect variability of the persistent westward-flowing current there. Oceanic currents usually have unsteady components with periods of a few days or longer.

Spatial differences also appeared in coherence

Figure 5-12. Low-pass (35-hour) filtered current (upper, cm/sec) and wind (lower, m/sec), from BC-2B 50 m. The wind data are from Fleet Numerical Weather Central estimates. Note the strong wind-driven pulse to the south during 19-23 January. Fig. 5-11 shows concurrent wind and ice conditions.

calculations. Vertical coherence was significant at tidal frequencies everywhere, but at low frequencies only in the coastal regime, perhaps because the water column there is homogeneous, unlike the other two regimes, which have strong stratification (during summer). This stratification may confine responses (e.g., to wind events) to only part of the water column. Horizontally, tidal coherences were large everywhere. Significant low-frequency squared coherences (~ 0.6 at 70 km separation) existed over the middle shelf during the summer of 1976, but were lower (~ 0.4) at similar distances between the middle and coastal regimes (i.e., across the inner front). Coherence squared was low in the outer regime and in the coastal regime, except at tidal frequencies (the number of records and their separations make this a tentative conclusion). The high coherence in the middle regime may be caused by the low background—the amount of low-frequency energy was small, and was mostly forced by wind events which have large spatial scales. Other processes may cause low-frequency motions in the outer and coastal regimes, and these competing processes can lower coherences calculated over long records.

At low frequencies the clear meteorologically driven pulses, as illustrated in Figs. 5-10 and 5-12, were observed only over the middle shelf. Some vestige of these was seen in coastal records, but not in the outer regime. It may be that the response to these wind events is so weak that it is swamped in the

regimes with more low-frequency energy (coastal and outer, see Table 5-3), but not in the middle regime.

These spatial changes in flow thus occur at many frequencies: tidal, subtidal, and mean (see preceding section). They are nearly congruent with hydrographic domains, so that the fronts delimiting these domains also bound flow regimes. Some of the relationship between the hydrographic domains and the flow regimes is clear. In the outer regime the mean flow agrees with the dynamic topography, as is true in the coastal regime and near the inner front. Variations in response to wind forcing are less clear, but are probably related to the changes in stratification and water depth.

SUMMARY

With a preliminary analysis, we have characterized the flow over the southeastern Bering Sea shelf. The spatial and temporal coverage of the measurements were perhaps two orders of magnitude better than previously available. We have calculated the frequency distribution of the flow, defined the separation of the shelf into three regimes, suggested annual variations, and identified meteorological forcing.

Flow above this shelf was mostly tidal: the scalar mean tidal speeds were about 20-25 cm/sec, increasing shoreward. About 80 percent of the tidal energy was semidiurnal (see Chapter 8, this volume). For individual records, up to 95 percent of the fluctuating energy was tidal, but both subtidal fluctuations and (record) means were significant. Differences in these low-frequency and mean flows defined distinct flow regimes nearly coincident with previously defined hydrographic domains (see preceding chapter).

In the outer regime, between the shelf break and the 100 m isobath, mean flow was westward at 1-5 cm/sec, agreeing with the dynamic topography. Energy at subtidal frequencies appeared unrelated to local meteorology, and may have resulted from oscillations of the Bering Slope Current and the associated front overlying the continental slope. Farther inshore, between the 100 m and 50 m isobaths, the middle regime had insignificant mean flow (although vigorous tidal currents exceeded 20 cm/sec). Near the inner front mean flows existed, however, at speeds of 1-5 cm/sec parallel to the 50 m isobath in a counterclockwise sense similar to calculations of geostrophic flow across the inner front. North of the Alaska Peninsula there appeared to be a net eastward flow during summer, but it seemed to be intermittent.

Ice cover and winds vary annually, and changes in flow were discernible over the seasons. In summer,

TABLE 5-7

Flow regimes

	Outer	Middle	Coastal
Fluctuating horizontal kinetic energy	60-70% tidal 1% inertial	90% tidal inertial (?)	80% tidal inertial (?)
Vertical coherence ¹	only tidal	only tidal	tidal and low frequency
Horizontal coherence ¹	only tidal	tidal and low frequency (C ² ~0.7 @60 km)	only tidal
Wind events	obscure	clearly present	present
Mean ³	1-5 cm/sec westward geostrophic balance	0.5 cm/sec random Note ²	1-5 cm/sec counter- clockwise geostrophic balance

¹The characteristics of coherence are particularly tentative. Missing records, improperly positioned instruments (e.g., both current meters on the same mooring in lower layer), and poor spatial coverage prevent strong conclusions. (C² is coherence squared.)

²In the vicinity of the inner front (near the 50 m isobath) speeds are 2-5 cm/sec and parallel the isobath (along-frontal) in a counterclockwise sense.

³Also see Figs. 5-9 and 5-10.

storms are less frequent and winds (in the absence of storms) are generally weaker (Chapter 3, this volume) than in winter. Currents directly driven by the wind were stronger and more common in winter than in summer within the coastal and middle regimes, but were not detectable in the outer regime in either season.

Further analysis of data already collected will make the description of shelf circulation more complete. Moreover, progress will be made in understanding the dynamics of the shelf circulation and its interactions with the variations in hydrographic structure across the shelf with the help of clues from more detailed examination of annual changes. Annually the mean wind stress, the wind stress variance, the stratification and its horizontal variation, the horizontal density (salinity) gradient, the freshwater discharge from rivers, and the extent of ice cover all undergo great changes. Understanding the effects of these on the circulation of this shelf will increase our

understanding of shelf dynamics in general, and help us to understand the ecosystem of the Bering shelf.

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Circulation and Hydrography of Norton Sound

6

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ABSTRACT

Norton Sound was two-layered vertically in temperature and salinity during summer, the eastern portion being more strongly layered than the western. Weak cyclonic mean flow in the eastern upper layer was not reflected in the lower layer, which was nearly stagnant and contained remnant cold, saline water formed locally during winter. Maintenance of this extreme layering in the eastern sound despite shallow depths (~15 m) was due to buoyancy input as solar heating and fresh water, sufficient to offset vertical mixing generated by tidal currents. In the western sound, coupled northerly upper- and lower-layer flows reflected a northward net flow over the shelf west of the sound. At times a strong baroclinic coastal flow occurred off Nome, and the central western sound was a locus for vertical mixing due to impingement of currents upon a shoal area. During winter, the sound approached vertical homogeneity due to vertical convection consequent to cooling and ice formation. Although it was possible to define mean flows, the currents were dominated by events which reflected regional wind and atmospheric pressure patterns in an as yet uncertain fashion. Such flow events may exert a primary control over such features as the Yukon River plume, which was never observed to enter the eastern sound even though the upper-layer salinity there suggests that Yukon water may have entered the sound prior to our observations.

INTRODUCTION

Observations of temperature, salinity, and currents have been obtained from Norton Sound, Alaska, as part of an investigation of transport processes on the Alaskan continental shelves. The general physical oceanographic conditions within Norton Sound, as deduced from observations made between 1976 and 1978, are addressed in this paper. Since analyses of

the observations are still in progress, this should be considered as an interim, working document rather than a set of firm, final conclusions.

Physical setting

Norton Sound is a shallow, high-latitude embayment extending eastward from the northern Bering Sea and forming an indentation in the central west coast of Alaska (Fig. 6-1). Its east-west length is about 220 km, and its width about 150 km. Depths vary from less than 10 m in the southern portion to more than 30 m in a trough-like feature which trends east-west in the nearshore region just south of Nome; average depth in the sound is about 20 m. Two promontories extend into the sound about two-thirds of the way toward its eastern end, Cape Darby from the north and Stuart Island from the south (Fig. 6-1).

Norton Sound is located in a region of extreme seasonal variability. During the approximately June-September summer the sound is ice free and air temperatures are well above freezing. The waters are exposed for 24 hours to daylight, though not necessarily direct sunlight, during part of this period, and generally to light and variable winds. By November, air temperatures drop well below freezing and ice formation has normally begun along the northern shore, with first ice typically forming on the surface in Norton Bay at the northeast corner of the sound.

Figure 6-1. Geographical location and bathymetry of the Norton Sound region.

Ice growth continues southward until, by mid-December, the entire sound is more or less covered. This ice cover, which usually persists until April or May, consists primarily of loose pack 0.5-1.0 m thick except for shorefast ice near shore and for some distance offshore in the region of the Yukon River Delta. Direct observations of the ice cover are limited, and much information on its distribution and extent has been obtained through the use of satellite data (Muench and Ahlnäs 1976, Ahlnäs and Wendler 1979).

During winter the ice cover would be expected to markedly reduce air-sea exchange of heat, moisture, and momentum. This would mitigate effects on the water of low air temperatures and winter storms.

Oceanographic background

Our existing knowledge of northern Bering Sea circulation before the present series of studies was summarized by Coachman et al. (1975). The regional circulation is dominated by a northward net water transport of about $1.5 \times 10^6 \text{ m}^3/\text{sec}$ over the shelf between Norton Sound and Siberia. [Editorial note: this value may be high. See next chapter. D.H.] More recently Muench et al. (1978), on the basis of near-bottom recorded current measurements, have reported northward net flow southeast of St. Lawrence Island and in Bering Strait from October 1975 through April 1977. They also noted that the currents were characterized by north-south flow events having

speeds of 50-100 cm/sec, fast relative to mean flow speed of about 15 cm/sec. These flow events had time scales of several days, or the same duration as meteorological events. Spatial variability of the northward flow and its interaction with the waters of Norton Sound remain uncertain.

Before this study, little oceanographic information was available from Norton Sound itself. Bottom sediment distributions within the sound suggest that mean circulation is cyclonic and penetrates to the easternmost sound (Drake et al. 1980), but flow rates have not been estimated. Cyclonic flow in western Norton Sound was depicted qualitatively in the compilation by Hughes et al. (1974) and mentioned as a probability by Coachman et al. (1975).

The Yukon River enters the Bering Sea in the southwest corner of Norton Sound (Fig. 6-1). A minor portion of its discharge flows directly into the southern sound, while the remainder enters the Bering Sea along the coast farther west. Both the volume and the pathway of Yukon water which enters the sound remain uncertain. River flow can be as high as about $3 \times 10^4 \text{ m}^3/\text{sec}$, as gauged at Pilot Station, some 60 km above the river mouth. Most of the annual river influx occurs between about April and November, and flow is an order of magnitude greater in midsummer than in midwinter.

FIELD PROGRAM

Oceanographic data were obtained from Norton Sound during 1976-78. Temperature and salinity data were acquired from vessels in summer 1976, 1977, and 1978 and through the ice from a helicopter in winter 1978. The periods of these cruises are summarized in Table 6-1. Twenty-four-hour time-series current observations were obtained from anchor stations in summer 1976, and instantaneous current profiles were obtained from anchor stations in summer 1977. Ancillary data obtained in summer 1977 included vertical profiles of transmissivity.

In addition to shipboard data, current data were obtained using taut-wire moorings during both winter and summer periods. Statistics for these moorings are summarized in Table 6-2.

Summer temperature and salinity data were obtained using a Plessey conductivity/temperature/depth (CTD) profiling unit with calibration samples on each cast. Winter data were obtained with a Plessey Model 9400 profiling unit operating through the ice from a helicopter. Temperature and salinity data are accurate to within 0.02 C and 0.02‰, respectively.

TABLE 6-1

Summary of cruises to Norton Sound

Vessel	Cruise ID	Dates	No. of Stations
<i>Discoverer</i>	RP4-D1-76B Leg V	26 Sept - 9 Oct 1976	33
<i>Sea Sounder</i>	None	8-12 July 1977	26
<i>Surveyor</i>	RP4-SU-77B Leg IV	11 Aug - 2 Sept 1977	55
<i>Discoverer</i>	RP4-D1-78B Leg I	10 July - 3 Aug 1978	56
<i>Discoverer</i>	RP4-D1-78B Leg III	10-29 Sept 1978	46
NOAA UH-1H	W-29	17 Feb - 5 March 1978	37

Currents were observed on the taut-wire moorings with Aanderaa Model RCM-4 current meters. Our discussion addresses the non-tidal, low-frequency flow components, as tides are treated in Chapter 8 of this volume. In order to remove tidal and higher-frequency components, the records were processed according to Charnell and Krancus (1976) and run through a 35-hour low-pass filter.

Anchored time-series stations were obtained using an Aanderaa current meter modified to operate through a deck readout unit. Speed and direction were measured hourly at 5-m intervals through the water column. Hourly CTD casts were also made at the time-series stations. Instantaneous current profiles were obtained using a Hydro Products current meter. Observations were not taken when the vessel exhibited yawing motions, easily detectable by rapid variations in heading.

In addition to the water column observations, wind speed and direction and air temperature were obtained from shipboard and were also recorded at the nearby weather station at Nome.

OBSERVATIONS

Distribution of density, temperature, and salinity

The waters of Norton Sound are characterized during the summer by two layers separated by a strong pycnocline. The upper layer is warmer and of

TABLE 6-2
Norton Sound taut-wire mooring summary

Mooring	Bottom D (m)	Meter D (m)	Mooring time period	Usable record L (days)	Latitude (° N)	Longitude (° W)	Vector mean speed	dir
NC-14	32	22	8/21/76–6/25/77	309	64° 21.6'	165° 21.6'	2.2 cm/sec	073° T
NC-15	19	15	8/21/76–1/30/77	131	64° 06.5'	165° 17.7'	5.9 cm/sec	018° T
NC-20	19	6	7/8/77–8/25/77	48	63° 59.7'	165° 29.4'	7.8 cm/sec	340° T
NC-20	19	14	7/8/77–8/25/77	None	63° 59.7'	165° 29.4'	malfunctioned	
NC-21	25	6	7/9/77–7/22/77	~13	64° 08.2'	163° 15.2'	malfunctioned*	
NC-21	25	20	7/9/77–8/26/77	48	64° 08.2'	163° 15.2'	0.1 cm/sec	198° T
NC-22	16	8	7/9/77–lost	None	63° 41.0'	163° 00.1'	not recovered	
LD-5	27	20	7/25/78–9/4/78	42	64° 08.3'	163° 00.2'	0.8 cm/sec	309° T

*Record became increasingly error-ridden with time, so no mean is presented here, but early part of record was useful for comparison purposes.

lower salinity, and hence is less dense than the lower layer. This layering was more pronounced and consistent in its characteristics in 1976 and 1977 (Figs. 6-2 and 6-3) than in 1978 (University of Washington data not shown). In 1976 and 1977 temperature and density differences between the layers were greater in the eastern than in the western sound. A near-surface lens of warm (>12 C), low-density ($<13 \sigma_t$ units) water in the southwestern sound was especially pronounced in July 1977 (station 23), and indicated the presence there of Yukon River water which is characterized by low salinity and corresponding low density. Concurrent temperature increases and density decreases with time during July-August 1977 were evident in both layers. The data in summer 1978 were insufficient to define spatial variations, but were adequate to detect higher bottom-layer temperatures in the eastern sound (~ 7 C, as opposed to 1-2 C during 1976 and 1977) than during the previous two seasons. Bottom-layer salinities were also lower in the eastern sound in 1978 than during 1976 or 1977 (about $32^{\circ}/\text{oo}$, as compared to $34^{\circ}/\text{oo}$).

The tendency for increased stratification in the eastern as compared to the western sound is exemplified by vertically averaged density gradients for July 1977 (Fig. 6-4). The single isolated high value off the

Yukon River was due to elevated near-surface temperatures. A pronounced region of minimum stratification was present in the central western sound. Closely spaced stations obtained through this region in August 1977 defined vertical density structure through the area of minimum stratification (Fig. 6-5). This feature coincided with a shallow rise (depths <20 m) in the bottom southeast of Nome (Fig. 6-1).

Though there was considerable year-to-year variation in detail, several features of the horizontal temperature and salinity distributions persisted. The surface and near-bottom temperature and salinity distributions adequately represent the upper and lower layers, respectively, and are used to depict these features (Figs. 6-6 and 6-7). The upper layer was dominated by an eastward-trending high-salinity, low-temperature tonguelike feature which originated in the northwestern portion of the sound. This feature was best defined in July 1977, and by August it extended east-southeastward the full length of the sound but was patchy in nature. In September 1976 the tongue was restricted to the region just off Nome, and in 1978 the data were inadequate to detect its presence or absence. Generally, lower-salinity, warmer water lay north and south of the tongue. The lowest salinities which were observed occurred off the Yukon River in July 1977. This may have been a

Figure 6-2. Vertical distributions of temperature along selected north-south sections across Norton Sound at three different times. Section locations are shown in Fig. 6-6.

Figure 6-3. Vertical distributions of density (σ_t) along selected north-south sections across Norton Sound at three different times. Section locations are shown in Fig. 6-6.

Figure 6-4. Horizontal distribution of the mean vertical density gradient in σ_t units/m for the upper 15 m of the water column during July 1977, obtained by vertically averaging, from the surface to 15 m, gradients computed for 1-m increments. Dotted line defining shoal area is schematic only.

sampling artifact, as measurements were obtained closer to shore at that time than during other cruises and probably penetrated farther into the Yukon River plume. The maximum observed temperatures ($>16^\circ\text{C}$) occurred in the northeastern sound during August 1977. The horizontal range of temperatures within the upper layer during September 1976 was only about 1°C , in sharp contrast to stronger gradients observed in the 1977 summer data.

The lower layer in the eastern and northeastern sound was characterized during summer 1976 and 1977 by particularly cold, saline water. The coldest ($<0^\circ\text{C}$), most saline ($>34^\circ/\text{oo}$) water observed was present in July 1977, whereas by August temperatures had increased to $2\text{--}3^\circ\text{C}$, salinities had decreased to $32\text{--}33^\circ/\text{oo}$ and the locus of maximum salinity had shifted somewhat westward. As for the surface layers, bottom-layer salinities were lowest off the Yukon River mouth.

A band of relatively low-salinity ($22\text{--}23^\circ/\text{oo}$) water paralleled the northern coast of the sound in September 1976 and August 1977 and appeared to be a westerly extension of warm, low-salinity water then occupying the upper layer in the northeastern sound.

Figure 6-5. Vertical distributions of density (σ_t) along selected sections normal to the coastline of northern Norton Sound at two different times. Section G-G' was occupied twice in rapid succession on the same day. Locations are shown in Fig. 6-7.

Figure 6-6. Horizontal distribution of upper- and lower-layer temperatures in Norton Sound at three different times. Location of sections in Figs. 6-2 and 6-3 are indicated.

Figure 6-7. Horizontal distribution of upper- and lower-layer salinity in Norton Sound at three different times. Locations of sections in Fig. 6-5 are indicated.

Station coverage during July 1977 and in 1978 was inadequate to determine whether or not the feature was present. Horizontal salinity gradients were stronger in both layers during 1977 than in 1976.

Other temperature-salinity features were evident only part of the time, but nevertheless can contribute to our understanding. In September 1976, temperature on the $\sigma_t = 21$ surface showed two tongues (Fig. 6-8): a warm (8.5-9.0 °C) easterly-directed tongue in the southern sound and a colder (7.0-8.5 °C) tongue extending westward in the northern part, neither of which was evident in summer 1977. Although distributions on isopycnal surfaces in shallow water like this, particularly near the top of the pycnocline, must be interpreted with caution, the distribution shown here agrees generally with the concept of a cyclonic circulation as discussed below. In September 1976 the near-bottom salinity distribution revealed two tongues of relatively high-salinity water ($>31\text{‰}$) penetrating the sound from the west, roughly coincident with the two troughs in bottom topography south of Nome (Fig. 6-7). These features were not evident in summer 1977. A westward baroclinic coastal flow was present off Nome during August 1976 (Fig. 6-5), and was also reflected in the time-series current observations at station 22 (Fig. 6-9). The same region during 1977 appeared to be the site of vertical mixing 10-15 km offshore rather than of such a baroclinic current.

Winter observations were obtained in February 1978, and the temperature and salinity distributions

at that time are summarized on vertical sections in Fig. 6-10. The entire water column was at the freezing point, or about -1.6 °C . Density differences were due to salinity differences, and were small with variations less than one σ_t unit. Since observations did not extend into the eastern portion of the sound because that region was ice free during the field work, conditions there remain uncertain, but they were probably similar to those observed in the western sound. The western sound was characterized by weak horizontal and vertical density gradients. Density was about 0.2-0.3 lower in the eastern than in the western portion, and lower by about the same amount near the bottom than near the surface. The two-layered structure which was present during summer had disappeared, its place having been taken by more or less uniform and weak vertical density stratification. Convective cooling and ice formation had created vertically uniform temperatures at the freezing point, but some vertical salinity, hence density, stratification remained.

CURRENT OBSERVATIONS

Current measurements were obtained from taut-wire moorings, as time series from anchored vessels and as instantaneous vertical profiles from anchored vessels. Statistics of the taut-wire mooring current measurements are given in Table 6-2.

Five time-series current profiles were obtained at anchor stations along a transect south of Nome during summer 1976 (Fig. 6-14). While these were of too short duration to use in estimating mean flow, four of the five (22, 24, 25, and 26) were long enough to allow averaging out of the tidal signal. Moreover, they provide the only vertical distributions of current obtained in the region south of Nome. They are summarized as vector-averaged currents at each depth and presented along with vector-averaged winds observed from the anchored vessel (Fig. 6-9). The highest speeds observed were about 50 cm/sec and occurred near the surface at Station 22. Speeds decreased with increasing depth at that location to about 10 cm/sec near the bottom, and the direction rotated from northwesterly near the surface to west-southwesterly near the bottom. Mean speeds at stations 24, 25, and 26 were highest—12-18 cm/sec near the surface—and decreased to about 5 cm/sec near the bottom. Flow directions at these locations were in the northwest quadrant throughout the water column. Short-term fluctuations, having time scales of a few hours, were superposed upon the mean flow at each location (Fig. 6-11). The surface wind was considerably steadier with time than the currents.

Figure 6-8. Distribution of temperature (°C) on $\sigma_t = 21$ in September 1976.

Figure 6-9. Vector-averaged currents and surface winds at the five time-series stations in western Norton Sound during September and October 1976. Mean current vectors are depth averaged. Station locations are shown in Fig. 6-14.

Two 44-day and one 14-day record were obtained from taut-wire moorings in summer 1977 (Table 6-2 and Fig. 6-12). An additional 45-day record, LD-5, was obtained during summer 1978 from nearly the same location as NC-21 off Cape Darby. Near-

bottom records from this location during both summers indicated flows of order 1 cm/sec or less. The summer 1977 near-surface record at the same location was cut short due to instrument malfunction, but indicated northwesterly flows of 10-15

Figure 6-10. Vertical distribution of salinity and density (σ_t) along two sections during winter 1978.

cm/sec. The summer 1977 near-surface record from NC-20 in the central western sound indicated north-northwesterly flow at 10-15 cm/sec.

Overwinter records were obtained from moorings NC-14 and NC-15 in the western sound. These were both near bottom and bracketed the period from late summer to mid or late winter 1977. They indicate a mean current which was northerly in the central sound and northwesterly along the northern shore, with speeds of about 5 cm/sec.

All of the moored current records revealed non-tidal speed fluctuations which were large relative to mean flow speeds and had time scales of several days (Fig. 6-12), including occasional flow reversals.

DISCUSSION

Winter hydrographic observations made through the ice have established that Norton Sound becomes vertically well mixed during winter. This is common

in high-latitude regions, as a consequence of vertical thermohaline convection due to surface cooling and ice formation. Observations during three summers reveal that well-mixed structure gives way to a regime which is strongly two-layered in temperature and salinity and hence also in density. Such a layered structure is generally typical of shallow oceanic regimes subject to tidal and wind mixing and buoyancy input. The upper layer is a consequence of the combined effects of wind mixing, freshwater influx, and solar warming, while the lower layer acquires vertical homogeneity through turbulence generated by currents at the bottom. The persistence of this structure in Norton Sound for several months each summer suggests that the time rate of change is small and that a small horizontal advection is balanced by other processes. The strong pycnocline between upper and lower layers inhibits vertical exchange of heat and salt. Our discussion examines these premises.

Figure 6-11. Hourly surface winds and currents at time-series station 25 in western Norton Sound during October 1976, illustrating short-term fluctuations.

Circulation

Horizontal property distributions tend to reflect circulation processes. During winter there was little horizontal variability; in summer the lowest temperatures and highest salinities in the upper layer occurred in the central western part of the sound. Elsewhere in the upper layer, temperature and salinity distributions, particularly in September-October 1976 and August 1977, indicate water mass continuity between the easternmost sound and a near-coastal band farther west along the northern coast. This structure suggests that a baroclinic coastal flow was transporting water from the eastern sound westward along the northern coast. The existence of this flow was substantiated both by subsurface density observations (Fig. 6-5) and current observa-

tions from station 22 (Fig. 6-9). The baroclinic feature was not present off Nome in August 1977, and hydrographic data obtained during July 1977 were inadequate to detect its presence or absence. Upper-layer current observations farther east off Cape Darby during July 1977 indicate that a westerly mean flow paralleled the coastline but was superposed upon a highly variable flow which included reversals. We conclude that westerly flow along the northern coast is a usual feature but may vary in intensity and extent. It was more pronounced and uniform during October 1976 than during August 1977.

Lower-layer temperature and salinity distributions differed from those in the upper layer, which suggests that advective processes were different in the two layers (Fig. 6-13). Temperature was lower and salinity higher in the eastern lower layer than in its

Figure 6-12. Low-pass filtered currents at 5 m depth at station NC-20 in Norton Sound, summer 1977; u and v are east and north components, respectively. Mooring locations are indicated in Fig. 6-14.

western part. This was most apparent in July 1977, less so in August 1977, and least obvious in September-October 1976. Deep temperatures in the eastern

sound increased during summer, while salinity decreased; during July-August 1977 the record from a current meter (NC-21) moored in the eastern lower layer indicated that these changes were nearly linear with time. The cold ($<0^\circ\text{C}$), saline ($>34^\circ\text{‰}$) deep water present in the eastern sound during summer 1976 and 1977 cannot have originated on the shelf to the west because water there was too warm and of too low salinity except in September 1976. This deep water can only have been a remnant of the preceding winter's convective layer, with elevated salinities resulting from salt exclusion as ice was formed. Observations during summer 1978 were sufficient to establish that this was not the case at that time as temperatures and salinities were similar in the eastern and western lower layers.

Sluggish circulation in the lower layer of the eastern sound was substantiated by observations. An advection rate along the lower boundary of the upper-lower layer interface can be estimated for September 1976 using temperature and salinity observations if we assume that the tongue-like low-temperature protrusion from the eastern sound (Fig. 6-8) reflected a steady-state balance between horizontal advection and lateral diffusion. A horizontal length scale of 100 km, applied to the empirical findings of Okubo and Ozmidov (1970), yields a lateral eddy conductivity of order $10^6 \text{ cm}^2/\text{sec}$.

Figure 6-13. Temperature-salinity curves from selected stations in eastern and western Norton Sound during the summers of 1976 and 1977. Temperature and salinities were in fact continuous from east to west across the sound, but stations from the central sound have been omitted in the interest of clarity.

Application of the graphical method described by Proudman (1953) to the geometry of the tongue leads then to estimates of a westward advection rate of less than 1 cm/sec along its axis. During July-August 1977 the observed lower-layer current speed at NC-21 off Cape Darby was 0.13 cm/sec, a comparable value. During summer 1978 a near-bottom mooring at the same location (LD-5) showed a net flow of less than 1 cm/sec to the northwest—again in agreement. The cause of this sluggish circulation in the eastern lower layer is uncertain, but it is probably the extreme layering which could decouple the lower layer from upper layer motions. It does not appear that basin configuration plays a major role, although the promontory extending from the southern shore formed by Stuart Island may be sufficient to deflect to the north whatever easterly flow exists along that shore, preventing it from entering the eastern sound. There exists, however, no sill between these promontories which might prevent interchange of deeper waters, and there are no observations to support this hypothesis. The case for decoupling of motions at the horizontal interface between upper and lower layers will be examined in more detail below.

While upper- and lower-layer circulation appeared decoupled in the eastern sound, time-series observations obtained from the vessel during September and October 1976 did not reveal such decoupling in the western sound. Rather, there was a monotonic decrease in current speed, along with a slight rotation of flow direction, with increasing depth (Fig. 6-9). The four stations with long enough records to allow at least a crude attempt at averaging out the tidal signal (stations 22, 24, 25, and 26) all indicated northwesterly surface flow which became westerly near the bottom. This was in rough agreement with the moored current record from July-August 1977 (NC-20) which indicated a northwesterly near-surface flow in the same region, with an overwinter record at NC-15 which showed northerly flow, and with a near-bottom record off Nome which indicated a weaker westerly flow during September 1976-March 1977. These current observations suggest that western sound circulation was dominated by flow which penetrated only slightly eastward into the sound and was driven primarily by the northward flow over the shelf between Norton Sound and St. Lawrence Island (Fig. 6-14). Low-frequency (of order two days and longer) current fluctuations observed at all of the moorings NC-14, NC-15, NC-20, NC-21, and LD-5 were a major characteristic of the flow; they suggest that the cyclonic circulation driven by northward regional flow may penetrate eastward into the sound to varying extents and may at times

Figure 6-14. Schematic representation of circulation and mixing in Norton Sound based on the summer 1976 and 1977 hydrographic and current data. Locations of time-series current and CTD stations occupied in September and October 1976 and taut-wire current moorings are shown.

form a cyclonic loop within the western sound. The occurrence of northwesterly flow fluctuations at NC-21 off Cape Darby suggests that at least the upper layer of the eastern sound responds to fluctuations in the western portion, although a coherency analysis of near-surface current records from stations NC-20 and NC-21 did not reveal a significant coherency between currents at the two locations. Penetration of flow eastward to Stuart Island would require, by continuity, increased westward flow off Cape Darby. That remnant water was not present in the lower layer in summer 1978 suggests also that advection had occurred in the eastern lower layer prior to our observations. Therefore, it appears that at times eastward flow may penetrate into the eastern sound.

The role of northward flow on the Bering Sea shelf in inducing a circulation within Norton Sound is uncertain. A probable mechanism invokes the high-latitude tendency for conservation of potential vorticity to constrain streamlines to parallel isobaths. The northward regional flow would tend to follow bottom contours into the sound and contribute to a cyclonic circulation there. The potential vorticity argument requires, however, that friction be neglected, an assumption of debatable validity in view of the shallow depths in the sound. Regardless of cause, easterly flow entering the sound along its southern

shore, curving cyclonically to the north, would then be deflected to the west at the northern coast to satisfy volume continuity constraints.

Subtidal flow pulses observed in the time-series current records are a dominant feature of regional flow. Their cause is uncertain, but the shallow depths and broad horizontal extent of the regional shelf suggest that winds and atmospheric pressure events may play major roles. Coachman et al. (1975) have suggested that flow through Bering Strait is characterized by pulsations that depend upon north-south atmospheric pressure differences across the strait. A detailed analysis of this problem is beyond the scope of this chapter, and is treated elsewhere in this volume (Chapter 7).

Maintenance of layering

The extremely sluggish circulation and consequent isolation of the lower layer in the eastern sound in the summers of 1976 and 1977 are of particular interest because of their basic scientific implications and because similar conditions may also exist in other high-latitude bodies of water subject to similar conditions—for example, Kotzebue Sound (Kinder et al. 1977). Retention of temperature and salinity characteristics by this water for four to five months in a total water depth of about 20 m, despite wind and tidal mixing and insolation, suggests: (1) horizontal advection of bottom water into the eastern sound was negligible; (2) insolation, particularly significant during early summer because of long daylight hours, had been unable to penetrate the bottom layer sufficiently to cause appreciable warming; and (3) vertical mixing through the pycnocline was extremely slight through the summer. Of these possibilities, the first has been shown by temperature, salinity, and current observations to be the case. The second may be examined summarily in the light of observations of transmissivity made by using a beam transmissometer in the eastern sound during summer 1977. In the northeastern sound off Norton Bay, transmissivity was of order 10 percent. Transmissivities were somewhat higher farther south, but were still too low to allow appreciable penetration of solar radiation below the upper layer. Transmissivities in the lower layer were uniformly of order 10 percent throughout the eastern sound. The net result was to effectively prevent solar radiation from penetrating to the lower layer.

The extreme layering in the eastern sound can be examined in terms of maintenance of the horizontal interface between layers. This interface is characterized by high vertical gradients; as high as 4 C/m in temperature and 1.5‰/m in salinity, with a cor-

responding Väisälä frequency of about 0.1/sec. Using the upper- and lower-layer current observations obtained from mooring NC-21 off Cape Darby, in conjunction with estimated mean vertical density distributions obtained at the beginning and end of the mooring period, a bulk interfacial Froude Number of about 10^{-1} can be estimated. This value suggests that little exchange due to turbulent mixing occurs across the interface. We may estimate a vertical mixing coefficient through the interface using a simple conservation of heat argument. Assume (1) that bottom water temperature at the end of each winter was near freezing (~ 1.7 C), as borne out by the winter 1978 observations, (2) that upper-layer temperature had reached 8 C by early August following a linear increase with time from early May when the ice first melted, and (3) that advective sources and sinks of heat can be neglected so that heat entering the lower layer all passes through the interface. The computed vertical mixing coefficient through the pycnocline was of order 10^{-2} cm²/sec for both summer 1976 and 1977. These values are small compared with those computed for other oceanic regions, but are reasonable for the extremely strong stratification observed.

Buoyancy input into the eastern sound appears to be critical to maintenance of the two-layered structure there. Comparison of our data with those presented by Coachman et al. (1975) from the Bering Strait suggests that the vertical density structure in the western sound is continuous with that on the Bering Sea shelf to the west. The enhanced summer 1976 and 1977 layering in the eastern sound can be accounted for by additional dilution of the upper layer by local freshwater influx and solar insolation. Assuming an initial salinity of 30‰ (see Fig. 6-10), about 10^6 m³ of fresh water are required for the observed dilution. This can be roughly accounted for by using a mean annual runoff of 30 cm extrapolated from Nome over a watershed area of about 2.5×10^5 km². The Yukon River is a second source of freshwater influx and during midsummer runoff of order 10^4 m³/sec could supply water sufficient to effect the observed dilution in about 10 days if the water were all to flow into the eastern sound. This amount of freshwater influx is equivalent to a buoyancy input of about 10^{-4} W m⁻². Assuming that solar insolation of 0.2 ly/min yields a buoyancy input to the upper layer of the same order, both insolation and freshwater influx are important to the buoyancy input. This compares with a bottom tidal dissipation of about 200×10^{-4} W m⁻² computed using the same method as Schumacher et al. (1979). In other shallow shelf regions, 1-2 percent of the tidal dissipa-

tion is sufficient to vertically mix the water column. Our estimates suggest that buoyancy input to eastern Norton Sound is sufficiently high that tidal dissipation is inadequate to vertically mix the water column. We conclude that the persistence of this extreme layering in the eastern sound is due to the roles of freshwater influx and solar insolation in generating sufficient buoyancy to overcome tidal mixing. Low net advection in the lower layer is then due to decoupling from the upper layer across the sharp interface maintained by buoyant forces.

Hydrocarbon observations

September 1976 observations of dissolved hydrocarbons provide additional evidence in support of our proposed general horizontal circulation scheme for Norton Sound. Although the observations included other hydrocarbons, we consider here distributions only of methane, ethane, and propane. Most methane found in shallow shelf waters arises from biochemical reactions occurring in anoxic near-surface sediments. A prominent local source of methane may be used as a short-term water mass tracer, since it appears to be quasi-conservative over time and space scales such as we are dealing with in Norton Sound.

The near-bottom methane distribution in Norton Sound suggests a dominant source (2,242 nl/l) at the eastern end of the basin, with a sluggish, ill-defined northwesterly drift (Fig. 6-15). Methane-deficient water (250 nl/l) was apparently entering from the south near the Yukon River Delta with relatively little transport eastward along the southern coast. The high methane concentration in the southeast corner of the sound suggests either an increased source at that location or an extremely sluggish circulation, or a combination of these.

Coincident with the methane observations, a near-bottom plume of ethane- and propane-rich water was detected about 40 km south of Nome near station 24 (Cline and Holmes 1977). The hydrocarbons appeared to originate from a point source and drift northward toward Nome, then northwest along the coast, in agreement with the circulation scheme deduced above.

The Yukon River plume

The fate of Yukon River water remains problematic, though it seems likely that it is dominated by flow events rather than mean flow. The temperature-salinity and hydrocarbon observations obtained during the summers of 1976 and 1977 yielded no evidence that Yukon River water was advected eastward into the eastern sound. Rather, it appeared that Stuart Island might have contributed to blocking the easterly flow along the southern coast (Figs. 6-5 and

Figure 6-15. Distribution of dissolved methane 5 m above the bottom, September 1976, in nl/l.

6-6). It is of course feasible that sporadic flow of Yukon water had occurred earlier during either or both seasons, and that the observed upper-layer dilution was due in part to remnants of past Yukon water inflows. This may have been true before July 1977, when particularly low ($<19^{\circ}/\text{oo}$) near-surface salinities were observed in the eastern sound (Fig. 6-7). The limited observations which indicated that the eastern basin had flushed out in summer 1978, resulting in deep-layer temperatures and salinities similar to those in the outer part of the sound, strengthen the supposition that appreciable eastward flow events may occur. The summer 1978 data were not however adequate to determine whether or not Yukon-diluted waters were in fact continuous between the eastern and western sound. A Yukon River origin for bottom sediments in the eastern sound suggests that Yukon water can enter there (Drake et al. 1980). The pulse-like events observed in the current records suggest, moreover, that water motion in the sound, particularly in the form of easterly migrations in the cyclonic flow, may be adequate to cause such additions. We have, however, no direct supporting evidence of such flow. R. Feely (PMEL unpublished data) has observed in data from summer 1979 a narrow coastal band of low-salinity surface water along the southern coast, which may be eastward-flowing Yukon water. However, no flow rates are available.

SUMMARY

Observations obtained from Norton Sound indicate

that in summer the regime was strongly two-layered in both temperature and salinity. The eastern and western sound were effectively separated into two distinct flow regimes at a line roughly coincident with the constriction formed by Cape Darby on the north and Stuart Island on the south. The eastern sound was more strongly layered than the western, and upper- and lower-layer flows were decoupled. The surface layer exhibited a weak tendency toward cyclonic flow, which was reflected both in the hydrographic observations and in moored current records. The eastern lower layer exhibited a near-zero mean flow allowing remnant cold, saline water from the previous winters' convective regimes to retain its identity in both 1976 and 1977 despite the shallow depth of the eastern basin. Sluggish lower-layer circulation was borne out by near-bottom moored measurements south of Cape Darby. Absence of the remnant water in summer 1978 suggested either that a flow event had flushed out the eastern basin or that less cold, saline water was produced in 1977-78.

Short-term time-series measurements in the western sound suggest that the upper and lower layers were not decoupled as in the eastern sound. These observations, in conjunction with moored current records and dissolved hydrocarbon distributions, support the concept of a northerly flow in the westernmost sound. The extent of this flow eastward into the sound may vary; it was apparently more intense in 1976, as evidenced by a strong westerly coastal current off Nome, than during 1977, when such a coastal feature was not observed. Large fluctuations relative to the mean flows, with time scales of several days, were observed at all moorings and suggest that instantaneous flow patterns in the sound may vary considerably.

The central western sound was a locus for vertical mixing as currents, primarily tidal, impinged upon a relatively shallow area. This was the only area where a breakdown in vertical layering was observed. We hypothesize that the relatively deep channels bracketing the shoal allowed occasional eastward flow of deeper, saline water that was then mixed upward leading to the observed higher local surface salinities.

At no time did we see flow of the Yukon River plume into eastern Norton Sound. It remains uncertain where the Yukon water goes: it appears likely that pulse-like flow events such as those observed at the current mooring sites may intermittently transport Yukon water into the eastern sound, or conversely, westward completely out of the sound. It may be entrained into the general northward flow west of the sound. Disposition of the Yukon River

plume must be considered an appreciable regional problem.

The near-bottom water in the eastern sound presents an interesting case. It is highly unusual for water at these depths (about 15 m) to retain its temperature-salinity identity for several months without undergoing vertical mixing due to winds or tides or sufficient lateral advection or diffusion to alter its heat or salt content appreciably. Maintenance of this regime is due to a large buoyancy input, in the form of insolation and fresh water, to the upper layer.

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Reevaluation of Water Transports in the Vicinity of Bering Strait

7

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INTRODUCTION

The general northward flow through Bering Strait transports water and associated properties from the Pacific Ocean into the Arctic Ocean. Although the strait's cross section is small ($85 \text{ km} \times 50 \text{ m}$) and the average transport is not large by oceanic standards ($\sim 1 \text{ Sv}$), the consequences of the flow reach well to the south into the Bering Sea and to the north into the Arctic Ocean. To the south, the effect of the Bering Strait flow is that of a continuous "leak" drawing off water from the Bering Sea across the large eastern continental shelf, where it significantly influences the flow field. To the north, Bering Strait water has been specifically traced to the North Pole (Coachman and Barnes 1961), and a recent analysis (Coachman 1978) suggests that the trans-ocean transport of the Arctic Ocean subsurface layer, to which the Bering Strait flow is a major contributor, may be much greater than previously supposed. Therefore Bering Strait water may have influence on the whole of the Arctic Ocean.

Even though the existence of northward flow has been known for over 300 years, and oceanographic studies have been made in the region using reliable methods since the 1920's (Sverdrup 1929), systematic studies of the flow employing direct current measurements on sections traversing the system between the Alaskan and Siberian boundaries began only in 1964 (Coachman and Aagaard 1966). Such measurements are necessary for determination of transport because of the spatial variability in the flow field. Results of these systematic studies through 1973 were summarized in Coachman et al. (1975). The available data came from 11 anchored current station sections across Bering Strait and another 10 sections north and south of the strait and extending entirely across the system between Alaska and Siberia. The latter can be used in assessment of

transport because runoff into the intermediate region is at least an order of magnitude less than the oceanic transport. Although a number of years are represented in the 21 sections, 19 were made during July and August, one in late September, and one in early October; hence, the results represent only summer conditions. The main findings were:

- (1) The north transport averaged close to 1.5 Sv .¹
- (2) There were temporal variations in transport equal to the mean flow on a time scale of days. The data included variations as large as 0.9 Sv in one day and 2.0 Sv in three days. There was one case of southerly transport (-0.2 Sv).
- (3) The force driving the transport was a sea-level slope generally down to the north, which remained unexplained. Variability in transport was due to a combination of wind and regional atmospheric pressure distribution. Local wind could modify transport by about 0.5 Sv for each dyne/cm^2 of sectional mean wind stress but could not account for the observed variations. The overall variability could be reasonably forecast by a linear correlation with surface atmospheric pressure at Nome with a phase lag of one day.
- (4) There was no evidence of an annual variation in transport. Among the many scattered current measurements available from the region, only two sets were made at times other than summer from which possible seasonal variations could be interpreted: one set from the Strait of Anadyr in February and another from the eastern channel of Bering Strait in April. These data provided no support for the seasonal variation in transport suggested by Maximov

¹ Throughout we use the convention that positive values are northerly and negative southerly.

(1945), Fedorova and Yankina (1964), and Antonov (1968).

One earlier report on transport through Bering Strait not discussed by Coachman et al. (1975) was that of Bloom (1964). He presented data on flow through the eastern channel for various times during the 1950's based mainly on electromagnetic measurements. Coachman and Aagaard (1966), in a critical review, concluded that the results could not be interpreted as transports because of uncertainties in the techniques and calibrations.

RECENT OBSERVATIONS

Since 1973 we have accumulated sufficient new data to allow a reassessment of the transport in the vicinity of Bering Strait, particularly of flow during seasons other than summer and of its variability. The new data include:

(1) Current records from meters moored at 11 locations from St. Lawrence Island to Cape Lisburne from September 1976 to June 1977 (Fig. 7-1), with the exception of mooring NC16, which was not deployed until the end of September. Each mooring, with subsurface flotation located well beneath the influence of ice, contained one Aanderaa current meter positioned 10 m above the bottom. The meters were set to record speed and direction at 40-minute intervals, and therefore nominally could record for one year; however, all the records ended for one reason or another between April and July. Concurrent data from all meters (except NC16) were obtained for the seven months September to March inclusive.

(2) Water level records obtained for the same period by Aanderaa pressure gauges mounted on the anchors of moorings NC7, NC10, and NC17, which recorded sea level at one-hour intervals.

(3) Sections of current meter stations between Alaska and Siberia from R/V *Moana Wave* during September 1976 (Fig. 7-1). These sections were the same as those reported in Coachman et al. (1975), and the measurements were taken with a deck read-out Aanderaa meter with the vessel at anchor.

(4) Five moorings with current meters in the area from outer Norton Sound to eastern St. Lawrence Island (in Fig. 7-1, the area from NC14 to NC16-17) for one and one-half months during summer 1978.

The main goals of these studies have been to clarify the temporal variations (periods of one day and longer) in regional transport, which have been shown to be very large on a short time scale, to secure the relationships between the variations and likely causal mechanisms (atmospheric phenomena), and to

Figure 7-1. Bering Strait and environs, showing location of long-term current meter moorings, sections of anchored current measurement stations, and weather stations.

assess the consequences of the variations for the ocean conditions of the northern Bering and southern Chukchi seas. Thus, among the new data the seven-month concurrent current records receive primary attention.

A reevaluation of the transport in the vicinity of Bering Strait using the new data requires an orderly series of analyses, to wit:

(1) Calibration of the single record from Bering Strait in terms of transport. This can be accomplished in two ways: by computing transports from the measurements of the six meters located along the Cape Lisburne section, which was a nearly closed section from Alaska to Siberia, and correlating the flow measured at NC10 with these; and correlating the transports from the 11 previous cross-sections with the flow at the position of meter NC10 in the section.

(2) Next, calibration of the flow measured in the St. Lawrence Island section (meters NC16, NC17) as transport. This is a more tenuous process, because although the flow passes through channels on both sides of St. Lawrence, the meters were placed only in the east channel. However, transport estimates are

possible using the two available measured sections transecting both channels (in 1968 and 1976).

(3) Next, testing of the transport variations as functions of various atmospheric parameters, including wind, pressure, and pressure gradients, which are available from daily surface pressure charts and weather stations located around the region.

(4) Finally, analysis of the response of the southern Chukchi and northern Bering seas to the observed variations in transport.

The present chapter, which should be considered part one of these analyses, concentrates on the Cape Lisburne section and the calibration of the flow measured in Bering Strait as transport, and concludes with a qualitative description of the relationship of transport to atmospheric pressure.

CAPE LISBURNE SECTION

The particulars of the Cape Lisburne moored section are given in Table 7-1, and the mooring locations are shown in Fig. 7-1. At each mooring the current meter, an Aanderaa RCM-4 cycling at 40-minute intervals, was suspended 10 m above the bottom. Except for mooring NC5, which was not recovered, complete monthly records were obtained from all the sites from September through March, and the present analysis concentrates on these seven months during which records from the entire section are available.

Table 7-2 gives the monthly and seven-month mean

velocities at the six instrumented sites. Four features seem particularly important:

(1) The long-term mean speeds are quite low, typically near 5 cm/sec, whereas we had *a priori* expected them to be 15-25 cm/sec even in the central Chukchi Sea (Coachman et al. 1975). These low mean speeds reflect not so much a quiescent regime as a regime highly variable in flow direction, thus resulting in a small vectorial mean. For example, every current meter recorded 40-minute mean speeds in excess of 40 cm/sec; at NC7 the maximum 40-minute mean speed was 68 cm/sec. However, as we shall see in detail in subsequent representations of the flow field, the directional variability is large. This is apparent even in the monthly means (Table 7-2), where from September to March every mooring except NC2 shows resultant monthly vectors in three separate quadrants.

(2) The monthly mean flow was most frequently toward the northwest or north at NC1-4, but during the seven-month comparison period had a southerly component nearly one-half the time at NC6 and 7. In fact, the strongest mean flows at the latter two moorings were southerly.

(3) There is some indication of a seasonal trend, with southerly flow most common in fall and winter. Thus, only two of 22 instrument-months showed a mean southerly flow component during March-September, while 10 of 30 did so during October-February.

(4) Table 7-2 also suggests some degree of hori-

TABLE 7-1. Cape Lisburne section, moorings NC1-NC7

Mooring	Latitude	Longitude	Depth of current meter	Series start	Series end
NC1	68° 15.4'N	172° 40.6'W	39 m	2200 GMT 25 August 1976	0520 GMT 17 June 1977
NC2	68° 29.7'	171° 55.3'	41	0000 26 August 1976	1520 4 April 1977
NC3	68° 44.2'	171° 06.2'	45	0220 26 August 1976	0020 7 August 1977
NC4	69° 00.7'	169° 59.2'	43	0600 25 August 1976	1600 6 May 1977
NC6	68° 57.2'	168° 18.6'	41	0220 25 August 1976	1140 20 April 1977
NC7	68° 55.2'	167° 21.3'	36	2140 24 August 1976	2340 27 June 1977

TABLE 7-2. Mean velocities

Mooring		Sept	Oct	Nov	Dec	Jan	Feb	Mar	7-mo. mean	April	May	June	July
NC1	cm/sec @ ° T	3.5 149°	1.2 057°	2.4 309°	6.8 327°	6.4 324°	3.8 353°	3.1 346°	2.6 335°	2.8 329°	4.4 339°	—	—
NC2	cm/sec @ ° T	4.0 292°	2.0 143°	1.3 280°	9.4 325°	8.2 316°	5.4 344°	3.7 326°	4.1 320°	—	—	—	—
NC3	cm/sec @ ° T	6.7 302°	2.5 175°	0.8 247°	6.1 333°	9.7 320°	5.6 322°	5.7 330°	4.5 317°	3.8 325°	5.0 340°	12.4 321°	8.1 307°
NC4	cm/sec @ ° T	7.4 334°	1.1 043°	2.0 132°	2.8 021°	6.8 344°	2.4 341°	4.8 008°	3.2 354°	4.4 335°	—	—	—
NC6	cm/sec @ ° T	8.0 359°	3.4 022°	1.7 061°	12.9 186°	5.5 230°	5.1 207°	3.0 012°	1.1 217°	—	—	—	—
NC7	cm/sec @ ° T	6.7 043°	6.4 076°	5.0 085°	17.8 182°	3.5 203°	4.9 223°	2.1 093°	3.1 145°	5.3 029°	4.3 035°	5.7 021° (27 days)	—

zontal coherence between adjacent instruments so that we might *a priori* have some expectation of meaningful transport calculations.

Figure 7-2 shows the daily mean vectors for NC1 and NC7 for September through March. The values have been smoothed by a 25-hour running mean.

Mooring NC1 has been taken as representative of the western and central parts of the Chukchi (NC1-NC4) and NC7 as representative of the easternmost Chukchi (NC6-NC7), where the strongest flows have previously been encountered (cf. Coachman et al. 1975, pp. 141-2). At NC1 the flow during the first

Figure 7-2. Daily mean current vectors at NC1 and NC7 for the period September through March, 1976-77, smoothed with a 25-hour running mean. The first day of each month is indicated by the appropriate capital letter.

three months of deployment was highly variable in direction, generally alternating between periods with northerly and southerly flow components. Peak speeds were typically 15-20 cm/sec and the variability time scale in the range of 3-10 days. In early December the speeds decreased to a slower level, typically 10 cm/sec or less, but much more nearly uniform in direction, generally NNW.

The record from NC7 is quite different. The flow there was considerably faster, commonly exceeding 20 cm/sec and not infrequently twice that or more. The first one-fourth of the record shows relatively few southerly flow events, but from late October on such events dominate much of the record. The time scale of these events is rather uneven. On the one hand, from early November to early December there was a remarkably regular pattern to the flow reversals, with a northerly set for two to three days followed by a southerly one for a like time. On the other hand, there were times of prolonged southerly flow, the most conspicuous beginning in mid-December and persisting for 23 days. Other flow reversal sequences are intermediate to these extremes. During the last month of the record (March) the speeds were decidedly lower, but the directional reversals continued.

Figures 7-3 to 7-9 show the temporal development of the current cross section month by month. The vertical axis is the time in days, and the six moorings are placed in their relative positions along the horizontal axis. The isotachs represent the vector daily mean flow component normal to the current-meter section (cf. Table 7-3 below). The hatched areas represent incidents of reversed flow, i.e., nominally southward.

The generally higher speeds on the eastern side,

TABLE 7-3. Transport section geometry

Mooring	Line segment length, l_i km	Mean depth, h_i m	Normal direction of l_i °T
NC1	58.8	50	318
NC2	41.9	53	322
NC3	48.5	54	325
NC4	57.3	53	346
NC6	53.0	50	005
NC7	57.5	42	006

Figure 7-3. Temporal development of flow in the Cape Lisburne section for September 1976. Ordinate is days, and current meters are placed in their relative geographic locations along the abscissa. Isotachs denote flow component normal to the section; hatched areas represent areas and times of reversed (nominally southward) flow.

Figure 7-4. Same as Fig. 7-3, for October 1976.

Figure 7-5. Same as Fig. 7-3, for November 1976.

the frequent flow reversals, particularly on the eastern side, and the coherence between adjacent mooring sites are all readily apparent. An interesting feature of these representations is the tonguelike character of the isotachs. That is, as a rule major

Figure 7-6. Same as Fig. 7-3, for December 1976.

Figure 7-7. Same as Fig. 7-3, for January 1977.

Figure 7-8. Same as Fig. 7-3, for February 1977.

flow events are most frequently connected with the ends of the section and show up at points increasingly further inside the section with reduced magnitudes. This is true of the strongest flow events, both northerly and southerly. Prime examples are 15-18 September (Fig. 7-3), 26-28 October (Fig. 7-4), 13-17 November (Fig. 7-5), and a whole series of southerly flow events along the eastern end of the section during December to February (Figs. 7-6 to 7-8). On occasion, a flow change at one end of the section

Figure 7-9. Same as Fig. 7-3, for March 1977.

appears appreciably later in the interior of the section. For example, southerly flow through the eastern end during 22 to 25 September appeared in the interior 28 to 30 September (Fig. 7-3). Furthermore, a *short* time lag of a day or so for flow events at the more interior stations is extremely common in these figures. The analysis and modeling of these flow events is outside the scope of the present description, but many of them are certainly suggestive of long waves propagating along the coasts (in this case, coasts with rather complicated geometry).

Figures 7-10 and 7-11 show the spectral distributions of energy for moorings NC1 and 7, again taken as representative of two different flow regimes. At NC1 the energy is relatively low, except for a strong

Figure 7-10. Spectral energy distribution for meter NC1, September-March 1976-77. Ninety-five percent confidence limits are indicated. "Rotated" indicates the spectrum is for the component of measured velocity normal to the section line, i.e., toward or away from Bering Strait (cf. Fig. 7-12).

Figure 7-11. Same as Fig. 7-10, for meter NC7.

signal at the M_2 tidal frequency. In particular, there is essentially no variance at time scales less than about three days. At lower frequencies the energy level increases gradually, with the bulk contained at frequencies less than 0.1 cycle per day.

At NC7 the flow is of an entirely different temporal nature. The M_2 signal is about one-third as large, but the low-frequency bands are far stronger everywhere below the diurnal. By far the largest energy concentration is the one with a period close to five days. We note also that the energy does not drop off at the lowest frequencies, as is the case for NC1.

In summary, then, the current records show a flow field which in the central and western Chukchi (NC1-4) is less energetic (except in the semidiurnal tidal band) and far less variable than in the eastern Chukchi (NC6 and 7). At the same time, the flow appears sufficiently coherent laterally to encourage transport calculations, with the expectation that there will be large variations at time scales longer than a few days.

Cape Lisburne transports

Figure 7-12 shows the transport section defined by the six moorings NC1-NC7, and Table 7-3 lists the geometric parameters.

The length of l_1 , i.e., the effective horizontal extent over which the current measured at NC1 is applied for transport purposes, was determined in the following way. Examination of *Brown Bear* (1960), *Northwind* (1963), and *Oshoro Maru* (1972) data (cf. Coachman et al. 1975 for data references) indicate that in the vicinity of l_1 , the influence of the south-

Figure 7-12. Geography of the Cape Lisburne transport section.

eastward-flowing Siberian Coastal Current extends seaward somewhat more than 90 km. The south-westward extension of line l_1 was therefore terminated at the indicated transition point to the cold, low-salinity Siberian coastal water. There is no guarantee that the Siberian Coastal Current will not from time to time cross line l_1 , and indeed the current and temperature record from NC1 indicates that on at least one occasion the direct influence of the coastal current may have extended as far seaward as the mooring position. On the other hand, the cross-correlations generally show significant length scales of 100 km or more, which is considerably greater than the distance from NC1 to the southwest end of l_1 , about 38 km. We can therefore expect the flow to be reasonably coherent over the distance l_1 , and the current record from NC1 to be indeed representative of the line l_1 . We return to the calculation of flow correlations later.

Line l_7 was extended eastward to the 30 m isobath, where the *Oshoro Maru* measurements still showed northeasterly flow; this is 6-7 km from shore. Since the water is quite shallow over the remaining distance, it is doubtful that much of the transport is missed. Lines l_2 - l_6 join points midway between each adjacent pair of mooring positions.

A more difficult problem is vertical extrapolation of the velocity from the instrument position 10 m above the bottom. Two shipboard current-meter sections have been made along the mooring line, one from the *Oshoro Maru* in July 1972 and the other from the *Moana Wave* in September 1976. For each of these sections we calculated the ratio between the mean cross-section velocity component in the layer deeper than 30 m and that in the water column taken as a whole. Omitting two apparently anomalous

stations of the total of 26, the mean ratio is 1.1: i.e., the deep flow was slightly higher than the mean over the water column. During the seven or eight winter months, when the water is vertically homogeneous, any baroclinic shear would of course be absent, and only the frictional shear (including that against the ice) would remain. In the absence of more extensive information, we therefore assume that the current 10 m above the bottom is representative of the water column as a whole. We further conjecture that the transport error introduced by this assumption is within 10 percent.

The temporal and spatial correlations of the flow field are critical for transport calculations. The spectral calculations (Figs. 7-10, 7-11) show relatively little energy at frequencies higher than the diurnal, except for the M_2 tide. This is in fact the case at all the mooring sites, as shown in Table 7-4, where the autocorrelations indicate a time scale of independence that exceeds two to three days. We should therefore expect that transport calculations done on a daily basis will cover all the important time scales. In Table 7-4 the autocorrelations have been calculated for four equal portions of the seven-month comparison period, and we note that the statistics vary not only from mooring to mooring, but also in time.

For the transport calculations to be meaningful, it is essential that the spacing of the current meters be less than the correlation length (cf. Fandry and Pillsbury 1979). The cross-correlations are shown in Table 7-5, again over 53-day intervals, as in Table 7-4. Table 7-5 provides a measure both of the mooring-to-mooring coherence and also of the correlation function from west to east (first column) and from east to west (last row). Except for the first portion of the

record, the flow field is well correlated from NC1 to NC4, a distance of 138 km, and from NC7 to NC4, a distance of 160 km. Throughout the record, adjacent moorings (diagonal) are correlated. The resulting daily transports (arithmetic estimates only, contrast Fandry and Pillsbury 1979) are presented in Fig. 7-13.

Bering Strait transports

The daily mean transports normal to the Cape Lisburne section are well correlated with the daily mean north velocities measured at NC10 ($r^2 = 0.81$). Transports through Bering Strait were calculated from the linear regression and are shown together with the Cape Lisburne transports in Fig. 7-13.

As a check, the eleven detailed Bering Strait sections were examined (10 of these are shown in Coachman et al. 1975, Fig. 48). From the isotachs of north-south flow, the velocities at the location of the NC10 current meter were estimated and related to the measured transport. The results, together with the linear regression of Cape Lisburne transport on Bering Strait north velocity, are depicted in Fig. 7-14; the good agreement suggests that the mean daily velocities measured at NC10 provide reasonable estimates of the transport through Bering Strait.

An obvious conclusion is that the whole southern Chukchi Sea, from Bering Strait to north of Cape Lisburne, is behaving as a coherent unit in response to the driving forces. There are two reasons why the coherence between Cape Lisburne transports and NC10 velocities is not even better:

(1) The shear in Bering Strait is not invariant. Thus, velocities measured at one point in the cross-section can only represent transport within some

TABLE 7-4. Number of days required for autocorrelation of component of velocity normal to sections to be statistically indistinguishable from zero

Mooring	Period 1 1 Sept. — 23 Oct.	Period 2 24 Oct. — 14 Dec.	Period 3 15 Dec. — 5 Feb.	Period 4 6 Feb. — 30 Mar.
NC1	5	4	6	3
NC2	6	4	4	4
NC3	5	4	3	3
NC4	5	4	10	4
NC6	3	3	8	3
NC7	3	3	4	3

Figure 7-13. Daily mean transports through the Cape Lisburne section and Bering Strait. Bering Strait transports were calculated from the correlation depicted in Fig. 7-14.

limits; Fig. 7-14 suggests the limits to be less than ± 1 Sv.

(2) The surface area of the southern Chukchi Sea, including Kotzebue Sound, between the measurement locations ($\sim 1.4 \times 10^5 \text{ km}^2$) rises and falls, partly in response to differential transports between the sections. (In this regard, inflow into Kotzebue Sound of the Noatak and Kobuk rivers is insignificant, the

TABLE 7-5. Cross-correlation of normal component of velocity, zero lag. For duration of numbered periods, see Table 7-4

	NC1	NC2	NC3	NC4	NC6
NC2	Per. 1 0.58 Per. 2 0.85 Per. 3 0.72 Per. 4 0.75				
NC3	Per. 1 * Per. 2 0.75 Per. 3 0.50 Per. 4 0.63	0.81 0.91 0.62 0.72			
NC4	Per. 1 -0.37 Per. 2 0.53 Per. 3 0.56 Per. 4 0.63		0.45 0.87 0.53 0.62		
NC6	Per. 1 * Per. 2 * Per. 3 0.30 Per. 4 0.48			0.27 0.80 0.69 0.45	
NC7	Per. 1 * Per. 2 * Per. 3 * Per. 4 0.45	* * 0.29 0.40	* 0.43 0.57 0.36	* 0.73 0.52 0.32	0.68 0.95 0.88 0.78

Asterisks denote velocity components uncorrelated at the 95 percent confidence level.

Figure 7-14. Correlation between mean daily Cape Lisburne transports and north velocity measured at NC10 in Bering Strait. Ninety-five percent confidence limits are indicated. Also shown are the transports from the previous 11 Bering Strait cross-sections of anchored stations related to the velocities at the location of NC10 in the section. (N.B. The correlation line is for Cape Lisburne transports vs. daily mean north component of velocity measured at NC10, $n = 211$, data not shown.)

maximum being $<10^4$ m³/sec.) Examination of the data in Fig. 7-13 shows that although sometimes the transports at the two locations were in phase, more frequently they were out of phase by a day in either direction, and this buffering effect of the southern Chukchi Sea reduces the coherence in transports between the two locations.

DISCUSSION

There are several surprises in the results, relative to the transport estimates of Coachman et al. (1975). These include the low mean values for longer-term flows (>1 month) and the extreme magnitudes of both the daily southerly transports and the accelerations.

The first surprise is the low value of long-term mean transport—over the seven months of record $\bar{T} = 0.3$ Sv, only about 20 percent of the annual mean value estimated by Coachman et al. (1975). This suggests that the previous estimate of mean annual transport of 1.5 Sv must be revised downward substantially, despite the fact that the mean daily transport values are of the same order as measured earlier and include even greater extreme values of both northerly and southerly transport. The cause of a reduced annual mean is the frequent incidence of southerly transport events during fall-winter, the season not covered previously by measurements. All previous data were from summer (July to early October), and of the 22 summer sections only two showed southerly transport (-0.1 , -0.2 Sv.). In contrast, Table 7-6 summarizes the incidence and duration of southerly flow events from the present mooring records.

TABLE 7-6. Incidence and duration of major southerly transport events, September 1976 through March 1977

Month	No. of events	Total duration, days
September	3	7
October	2	11
November	5	9
December	4	12
January	3½	8
February	1½	8
March	2	6

A relatively high incidence of southerly transport events in fall agrees with Bloom's (1964) measurements, made with electrodes spaced over a distance of 7.5 km west from Cape Prince of Wales. Although

the potential measurements cannot be converted to transports, they suggest possible southerly flow events and their duration. Fig. 7-15 compares such suggestions with measured southerly flow events during 1976 and 1977. In both years the period September through December showed significant southerly flow, but January through March 1976-77 showed definitely more southerly flow events than the similar period in 1957.

Figure 7-15. Incidence and duration of southerly flow events in Bering Strait, from the 1976-77 measurements and from the electrical potential measurements of Bloom (1964) for 1956-57.

We can construct a revised estimate of mean annual transport. The Cape Lisburne transports were extrapolated to the period April-June by linear regression of the September-March transports on the mean speeds (normal to the station line) measured at NC3 and NC7, which had records extending through June. The correlation was $r^2 = 0.8$; the regression equation gives for April and May, $T = 0.50$ Sv, and for June, $T = 0.84$ Sv. If we now assume that 1976-77 was a representative year, and that the previously determined average of 1.5 Sv applies to July and August, the mean annual transport would be 0.6 Sv.

Soviet oceanographers have published many estimates of the long-term transports, but nowhere can we find a description of the data or observational methods on which their results are based. Table 7-7 contrasts our results with the values reported by Fedorova and Yankina (1964), who included results from eight other studies. There appears to be an annual cycle of transport magnitude, from low mean monthly values in winter (0-0.5 Sv) to high values in August (1.5-2.0 Sv). But we do not believe the cycle should be taken too literally, because in any one month the mean transport can vary significantly, depending on the incidence and duration of southerly flow events that affect the longer-term mean transport values. These monthly values, in turn, will

TABLE 7-7. Monthly and annual Bering Strait transports in Sv

	Soviet results ^a		Previously measured cross sections		
	1941-1961	Range	1976-1977	Averages	No. of sections
Jan	.60	.60-.86	.54		
Feb	.50	.50-.94	.24		
Mar	.46	.46-.70	.33		
Apr	.59	.58-.80	.50 ^b		
May	.86	.76-1.00	.50 ^b		
June	1.24	.94-1.24	.84 ^b		
July	1.56	1.01-1.77		1.0	(14)
Aug	1.62	1.13-2.25		1.7	(5)
Sept	1.35	0.38-2.37	.46	1.3	(2)
Oct	.99	0.19-2.13	.03	1.3	(1)
Nov	.80	0.38-1.66	.32		
Dec	.68	0.56-1.20	.13		
Mean Annual	.95	0.31-1.42	0.6		

^afrom Fedorova and Yankina, 1964^bextrapolated

probably depend primarily on the pattern of atmospheric conditions which happens to prevail during the period. In any event, our monthly mean values appear to be, if anything, a little lower than the compiled Soviet results. Fedorova and Yankina also reported annual means for 1952 through 1961 which ranged between 0.85 Sv and 1.08 Sv to the north. Comparison of the Soviet results with those from 1976-77 and the earlier measurements suggests that the 1976-77 annual estimate of 0.6 Sv is probably a minimum, and also that there undoubtedly are interannual variations. Under these considerations, our present best estimate is 0.8 ± 0.2 Sv for the annual mean.

A second major surprise in the 1976-77 records is the high values of daily mean transport. The range of values, 3.1 Sv northerly to -5.1 Sv southerly, extends the earlier range estimates (Coachman et al. 1975) considerably, particularly on the southward end. In the earlier results, the highest value was 2.2 Sv northerly; in the new results only three values from Bering Strait and four from Cape Lisburne exceed this. However, in southerly flow the earlier extreme value was only -0.2 Sv. In 1976-77, 26 daily values exceeded -1 Sv, of which eight were associated with

the major flow reversal of Oct. 24 through Nov. 1 (Fig. 7-13).

In the first attempt to relate mean daily transports to atmospheric conditions, Coachman et al. (1975) noted that low northward transport seemed to be correlated with low atmospheric pressure at Nome with a phase lag of ~ 1 day, and vice versa. They also noted that the best correlation for the 21 cases ($r^2 = 0.74$) included both atmospheric pressure and a northwest-southeast pressure difference across the system (pressure at Nome minus pressure at Cape Schmidt).

As the first step in interpreting the relationship between transport variations and atmospheric conditions for the seven months of record, daily mean surface pressures were obtained from Nome (P_n) and Kotzebue (P_k) on the eastern side of the system and Provideniya Bukhta (P_p) and Cape Serdtse-Kamen (P_{sk}) on the western side, the latter two stations being located south and north of Bering Strait. (N.B. Apparently Cape Schmidt is no longer a reporting station; however, Cape Serdtse-Kamen, to the northwest of Bering Strait, lies due west of Kotzebue with the same east-west separation from it as Nome from Provideniya Bukhta, see Fig. 7-1.) All possible correlations were investigated, with the following results:

(1) The surface atmospheric pressure pairs on the eastern (P_n and P_k) and western (P_p and P_{sk}) sides of the system were each highly correlated, but there were significant day-to-day E-W pressure differences and variations. The east-west pressure differences south ($P_n - P_p$) and north ($P_k - P_{sk}$) of the strait were well correlated ($r^2 = 0.80$).

(2) The Bering Strait transports were well correlated with the east-west pressure differences. The best correlation ($r^2 = 0.70$) was a two-component correlation using both ($P_n - P_p$) and ($P_k - P_{sk}$), but the correlation was essentially the same ($r^2 = 0.68$) using only ($P_n - P_p$). The relationship was nearly in phase. The correlation between transport and ($P_n - P_p$) the same day was $r^2 = 0.62$, but only 0.53 for ($P_n - P_p$) one day earlier; the best result was obtained by averaging ($P_n - P_p$) for the same day with one day earlier. The resulting equation is

$$T = 1.10 + 0.20 (\overline{P_n - P_p}),$$

where T is in Sv and $(\overline{P_n - P_p})$ in mb. The daily mean Bering Strait transports and east-west pressure differences for the seven-month records are shown in Fig. 7-16.

(3) The transports were not correlated with any of the individual station pressures or other combinations of pressures, regardless of lag. This result is

contrary to the earlier suggestion (Coachman et al. 1975) that the surface pressure at Nome one day earlier could be used as an index of the transport. The mean atmospheric pressure at Nome for September-March was about 10 mb lower than that for the earlier transport calculations, and strong south transport did not correlate with low pressure at Nome; about one-half the incidences of south transport occurred when P_n was greater than 1,000 mb. Apparently, there were too few data in the previous analysis, and furthermore those cases were from summer, when there is more sustained northerly transport and atmospheric pressure at Nome tends to be higher.

To explore the relationship between atmospheric weather patterns and transport variations, we examined the mean daily surface atmospheric pressure charts for the northern hemisphere, using the chart for 0000 Z to represent the previous day (the longitude of Bering Strait, 165°W , is time zone +11). The water level records from pressure gauges mounted on moorings NC7 and NC17 (Fig. 7-1) provide additional data useful for interpreting the transport accelerations. Because the meters were at different depths, the measurements can be used only to suggest the relative changes in north-south water level along the axis of the system. The mean daily values are shown in Fig. 7-16 as anomalies of water level from their mean values over the seven months. Also plotted is the difference between water level anomalies at the two meters, with positive values corresponding to the

sea level at St. Lawrence being higher than normal, relative to sea level at Cape Lisburne; that is, sea level sloped more strongly down to the north.

First, examination was made of pressure patterns surrounding all the major south flow events, e.g., 16 September, 10-11 October, 26-28 October, 22-23 November, et seq. In every case the large-scale atmospheric pressure patterns were the same. One day before a peak in southerly flow, a strong low-pressure system was centered some distance to the southeast of Bering Strait, in the area of Bristol Bay, Kodiak, Anchorage, and the northern Gulf of Alaska. At the same time the Siberian high was centered some distance west or west-northwest of the strait. The isobars signifying the strongest pressure gradient between pressure centers were located precisely over the Bering Strait region and, most significantly, they had a nearly north-south orientation which extended from over the Chukchi Sea south into the central Bering Sea—completely across the northern Bering Sea shelf. If the north-south orientation of the isobars did not extend totally across the northern shelf, or if the isobars were oriented northeast-southwest (the most typical configuration), strong southerly flow events did not occur.

Fig. 7-17, surface pressure charts for 20-23 November, illustrates a typical case. A low over the Gulf of Alaska on 20 November moved over Kodiak on 21 November and deepened. At the same time, the center of Siberian high pressure moved west. Thus, on 21 and 22 November the strongest pressure

Figure 7-16. Daily mean values of transport through Bering Strait, E-W surface atmospheric pressure differences across the region ($P_n - P_p$) is south of Bering Strait and ($P_k - P_{sk}$) north, cf. Fig. 7-1), water levels measured at St. Lawrence Is. (NC17) and Cape Lisburne (NC7, cf. Fig. 7-1) plotted as anomalies from seven-month mean values, and the differences in water-level anomalies (south minus north).

Figure 7-17. Charts of surface atmospheric pressure for 0000 Z 21-24 Nov. 1976, taken to represent midday conditions the day before. The chart covers the area from the north Pacific to the Arctic Ocean, with Alaska on the right and Siberia on the left.

gradients were directly over the Bering Strait region and the isobars had an extended north-south orientation. The situation held over 22 November, although the pressure gradient was less as the low began filling, and by 23 November these conditions had dissipated. Southerly transports of -1.9 Sv and -1.8 Sv were recorded on 22 and 23 November (Fig. 7-16).

The mechanism which drives major south flow events now seems clear. Strong north winds must develop over the whole northern Bering Sea, not just over the immediate region of Bering Strait. Large-scale, strong atmospheric pressure cells are required, a low well to the southeast and a high well to the west.

The strong northerly winds generated thereby move water southward off the entire northern Bering Sea shelf. Removal of sufficient water off the northern shelf generates a sea-level slope down to the south (sea-level slope has been shown to be the major force driving transport through the strait (Coachman et al. 1975)). This, together with the strong north winds caused by the east-west atmospheric pressure gradient aligned over the system, drives enhanced southerly transport. It apparently requires about one day for development of these conditions, so that maximum south transport occurs the following day. Because the system behaves to a marked degree as a coherent

unit, water levels at both St. Lawrence and Cape Lisburne fall together and are nearly in phase with the transport (cf. Fig. 7-16).

This mechanism also accounts for the nearly in-phase behavior of transport and $(P_n - P_p)$. Since it is the removal of water from the whole northern shelf area (largely located south of St. Lawrence Island) that is responsible for major south transport, the strong east-west pressure difference associated with the proper atmospheric condition does not have to lead the transport if the weather system develops from the south. Water removal from the northern shelf can begin before the pressure gradient indicating a strong north wind is registered at Nome and Provideniya Bukhta, located north of St. Lawrence and closer to Bering Strait.

Northward transport stands in contrast to the southerly transport events. Periods of northerly flow tend to be more persistent and not so great in magnitude, nor do they show the marked episodic character of the southerly flows (Fig. 7-16). The greater persistence of northerly flow must reflect the basic driving force, a higher sea level in the Bering Sea than in the Arctic Ocean (Coachman et al. 1975), which still remains unexplained. There were, however, a number of relatively rapid northward accelerations of transport during the seven months of record, which appear to have two basic causes:

(1) After strong south transport events, rapid accelerations commonly occur which can be thought of as compensatory accelerations. When atmospheric conditions causing the southerly transport event dissipate, water is not being removed from the northern shelf, but there is still large southerly transport in the system. Water "piles up" in the region around St. Lawrence Island and Norton Sound, a condition reflected by a strong positive difference in water level anomalies, and following this by about one day a strong northward acceleration occurs. For example, on 23 November the atmospheric conditions driving the southerly flow of 22-23 November had dissipated (Fig. 7-17) and on 23-24 November the St. Lawrence water level anomaly rose significantly above that at Cape Lisburne (Fig. 7-16). Bering Strait transport changed from -1.8 Sv on the 23rd to $+1.1$ Sv on the 25th.

(2) Occasionally, major northward accelerations appear to be, at least in part, directly driven by atmospheric conditions. Specifically, these are a strong pressure low centered in the western Bering Sea, southwest of Bering Strait, or a deep trough from the central Aleutians toward the northwest, so that the isobars in the strong pressure gradient are directed northward from the central Bering Sea along

the axis of the system. This configuration creates strong southerly winds which can move water from the central Bering Sea onto the northern Bering Sea shelf, raising the water level in the vicinity of St. Lawrence Island and enhancing the sea-level slope down to the north. The best example during the seven months is depicted in Fig. 7-18. The proper isobar configuration was present on 5 January, while northerly transport accelerated to 2.5 and 2.6 Sv on 6 and 7 January, respectively (Fig. 7-16). In this case northerly transport was probably enhanced by northward extension of the east-west pressure gradient past Cape Lisburne, indicating southerly winds over the southern Chukchi Sea. These could have abetted northerly flow by removing water from the downstream side of the strait. Water levels at both St. Lawrence Island and Cape Lisburne rose considerably as a consequence of water being advected into the system from the south, and the increases were in phase with the transport, reflecting the coherent behavior of the system.

5 JAN 1977

Figure 7-18. Same as Fig. 7-17, for 5 Jan. 1977.

The prolonged southerly flow event of 24 October through 1 November (Fig. 7-16) illustrates the proposed set of mechanisms well. Development of the proper atmospheric conditions for southerly flow was first noticeable on the chart for 24 October. The low was centered over Anchorage, and the maximum pressure gradient west of the low extended from Pt. Barrow to the Pribilof Islands. It reached an extreme value on 16 October (low = 968 mb, high = 1041 mb), followed by the maximum measured southerly transport of -4.5 Sv on 27 October. The low remained stationary and had partially filled by the 29th, but was replaced by another intense low (956

mb) on the 30th. Thus, the northward acceleration of 28 and 29 October, caused by an enhanced sea surface slope down to the north following the partial filling of the first low (strong positive water level anomaly difference on the 27th: Fig. 7-16) was arrested through 1 November by the new intense low forming on 30 October. The pattern of closely spaced isobars extending north-south on 31 October had disappeared on 1 November, after which there was another northward acceleration.

SUMMARY AND CONCLUSIONS

New data on the flow regime in the vicinity of Bering Strait obtained since the comprehensive analysis of Coachman et al. (1975) include seven-month (September 1976-March 1977) current-meter records along a nearly closed section from Cape Lisburne to Siberia and from one instrument in Bering Strait. Daily mean currents normal to the Cape Lisburne section showed that the flow field of the southern Chukchi Sea divides into two regimes. In the western half currents were weak (<20 cm/sec). Except for a strong M_2 tidal signal, the bulk of the energy was contained at frequencies less than 0.1 cycles per day. In the eastern half, the currents were much swifter (up to 68 cm/sec), but with less energy at the M_2 tidal frequencies. At subtidal frequencies, the energy levels were far higher than to the west, with the largest energy concentration centered near the five-day time scale. Unlike what the records show for the western half, the energy did not drop off at the lowest definable frequencies. Throughout the section there were numerous current reversals (to the south) which were concentrated in the fall and early winter. These southerly flow events tended to begin at the ends of the section and propagate into the interior of the sea and are suggestive of long waves propagating along the coasts.

Daily mean transports across the Cape Lisburne section were well correlated ($r^2 = 0.81$) with mean daily north velocities measured in Bering Strait. Thus the whole southern Chukchi Sea from Bering Strait north past Cape Lisburne is responding as a coherent unit to the driving forces. These data, taken together with the previous 22 cross sections of measured flow, allow calibration of the Bering Strait measurements as transports. Daily mean values ranged from +3.1 Sv (north) to -5.0 Sv (south), extending the range of previously measured values considerably, particularly toward the south. There were very large daily accelerations in transport, in six instances exceeding 2 Sv/day.

Longer-term mean transport values (monthly, annually) are determined by the incidence and

duration of southerly flow events. Such events were most frequent in fall and winter 1976-77, corresponding with the results from electric-potential measurements in Bering Strait over the year 1956-57 (Bloom 1964). More frequent and prolonged southerly flow events in fall-winter probably give rise to an annual cycle in Bering Strait transport, with monthly mean values varying from 0-0.5 Sv northerly in winter to 1.0-2.0 Sv northerly in summer. The mean annual transport appears to be $\sim 0.8 \pm 0.2$ Sv.

Analysis of the relationship between transport and the atmospheric pressure field suggests a good correlation ($r^2 = 0.7$) between transport and the east-west pressure difference across the system. When pressure on the east side (Nome, Kotzebue) is greater than on the west side (Provideniya Bukhta, Cape Serdtse-Kamen), the regional air flow is towards the north, as are the water transports, and vice versa. Transports are not correlated with atmospheric pressure at Nome one day before, as suggested earlier.

Southerly transport through the strait has a marked episodic character, and events are caused by particular large-scale atmospheric conditions. When a strong east-west pressure gradient lies over the strait and extends in a north-south direction from the Chukchi Sea to the central Bering Sea, entirely crossing the northern Bering Sea shelf, extensive northerly winds move water southward off the shelf. This produces a sea-level slope down to the south which, together with the northerly winds, drives southward transport. The atmospheric pressure pattern causing this condition is always a strong low located a considerable distance southeast of the strait (e.g., over Kodiak) together with the Siberian high centered some distance west of the strait. Peak southerly transport follows development of such a condition by about one day.

Northward transport is more persistent, does not show the episodic character of southerly transport, and the transport values are lower than those of peak southerly flow. The basic driving force is an as yet unexplained higher mean sea level in the Bering Sea than in the Arctic Ocean. Strong northward accelerations have two causes. Most common is a compensatory acceleration following a southerly transport event; dissipation of the driving force for the southerly flow event, but continued southerly transport through Bering Strait, raises water level in the vicinity of St. Lawrence Island/Norton Sound, resulting in an enhanced sea-level slope down to the north driving northerly acceleration. The second mechanism, which rarely occurred during the seven months of record, was the development of a strong low-pressure center in the western Bering Sea, south

and west of the strait, with isobars in the strong pressure gradient directed northward from the central Bering Sea to the strait. The extensive strong south-

erly winds abetted northerly accelerations by moving water into the northern Bering Sea shelf, thereby enhancing the sea-level slope down to the north.

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Tides of the Eastern Bering Sea Shelf

8

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ABSTRACT

The acquisition of a substantial amount of pressure-gauge and current-meter data on the Bering Sea shelf has permitted a much more accurate description of the tides than has previously been possible. Cotidal charts are presented for the M_2 and, for the first time, the N_2 , K_1 , and O_1 constituents, and tidal current ellipse charts for M_2 and K_1 . S_2 , normally the second largest semidiurnal constituent, has not been included because it is anomalously small in the Bering Sea. The tide enters the Bering Sea through the central and western Aleutian Island passes and progresses as a free wave to the shelf. Largest tidal amplitudes are found over the southeastern shelf region, especially along the Alaska Peninsula and interior Bristol Bay. Each semidiurnal tide propagates as a Kelvin wave along the Alaska Peninsula but appears to be converted on reflection in interior Bristol Bay to a Sverdrup wave. A standing Sverdrup (Poincaré) wave resulting from cooscillation in Kuskokwim Bay is evident on the outer shelf. The semidiurnal tides are small in Norton Sound where there is an amphidrome. The diurnal tides, which can have only Kelvin wave dynamics, cooscillate between the deep basin and the shelf. Amphidromes are found between Nunivak Island and the Pribilof Islands, and west of Norton Sound. Throughout most of the shelf the tide is of the mixed, predominantly semidiurnal type; however, the diurnal tide dominates in Norton Sound.

Tidal models by Sünderman (1977) (a vertically integrated M_2 model of the entire Bering Sea) and by Liu and Leendertse (1978, 1979) (a three-dimensional model of the southeastern shelf incorporating the diurnal and semidiurnal tides) are discussed. Good qualitative agreement is found between the models and observations.

INTRODUCTION

As with most continental shelves, the tides and tidal currents on the eastern Bering Sea shelf play important roles in such oceanographic processes as the maintenance of the density structure, sediment resuspension and transport, and the distributions of benthic and intertidal organisms. A knowledge of the tides and tidal currents is therefore necessary in order to understand the region's oceanography. The tides of the Bering Sea have been of interest to physical oceanographers and astronomers for a long time (e.g., Jeffreys 1921, Munk and MacDonald 1960, Cartwright 1979). This interest has been based on the premise that the vast continental shelves of the Bering Sea (Fig. 8-1), with their proximity to the Pacific Ocean, act as a major sink of the world's tidal energy. Yet many aspects of the tides and tidal currents in the Bering Sea have remained unknown because inadequate data made it impossible to draw definitive cotidal charts or to obtain reliable boundary conditions for numerical models. Fortunately, in recent years a large number of pressure-gage and current-meter observations have been made on the eastern Bering Sea continental shelf, and the new data make possible a more detailed description of the tides in the eastern Bering Sea.

Figure 8-1. Bathymetric chart of the Bering Sea.

Concurrent with the recent field work has been the development of several numerical models for tides in part or all of the Bering Sea. One of these, the Liu and Leendertse (1978, 1979) model of Bristol Bay, will be described in detail. The other models are discussed briefly for completeness and to show the reader the scope of current theoretical work. The field work is also continuing. In a sense then, this chapter is a progress report as well as a review of past work and a description of new results.

The tides that we shall be concerned with are the principal tidal constituents N_2 and M_2 in the semi-

diurnal band (periods of about 0.5 days) and O_1 and K_1 in the diurnal band (periods of about 1.0 day). Ordinarily, the principal solar semidiurnal constituent S_2 would be included in the discussion. However, S_2 is anomalously small throughout the Bering Sea, possibly because it has small amplitudes in the adjacent North Pacific Ocean. Whatever the reason, no consistent distribution for S_2 appears in the field data above the background noise level. The complicated distributions of semidiurnal and diurnal tides in the Bering Sea produce a rich variety of tidal types, ranging from fully semidiurnal in some regions to

fully diurnal in others. Sample time series will be shown later in the chapter to illustrate the tidal types. Probably the most important figures are the cotidal charts for the four principal constituents, because from these charts it is possible to infer much about the dynamics of the tides and to obtain harmonic constants for tidal predictions. Empirical cotidal charts derived from recent data and theoretical cotidal charts from models will be presented.

The discussion of tidal currents in the Bering Sea will be presented with less certainty than that of the tides, because the few early measurements of tidal currents were of short duration and mostly limited to harbors and nearshore regions, and the recent observations from offshore current meters occasionally suffered from errors due to biological fouling and effects of wave motion. Still, with careful study and editing of the recent data reasonably accurate tidal current harmonic constants can be obtained for most areas. These results help define the dynamics controlling the tidal motion.

The setting for the tidal and tidal current distributions can be obtained from previous work (Harris 1904, Leonov 1960, Office of Climatology and Oceanographic Analysis Division 1961, Defant 1961, Coachman et al. 1975) which for the most part describes the semidiurnal tide. The tide wave enters the Bering Sea as a progressive wave from the North Pacific Ocean, mainly through the central and western passages of the Aleutian-Komandorski Islands. The Arctic Ocean is a minor secondary source of tides which propagate southward into the north Bering Sea where they complicate the tidal distributions. The North Pacific and Arctic Oceans are also sinks of tidal energy for tides propagating out of the Bering Sea. Tides in the Bering Sea are considered to be the result of cooscillation with large oceans. Once inside the Bering Sea, each tidal constituent propagates as a free wave subject to the Coriolis effect and bottom friction.

The tide wave propagates rapidly across the deep western basin. Part of it then propagates onto the southeast Bering shelf where large amplitudes are found along the Alaska Peninsula and in Kvichak and Kuskokwim Bays. Another part propagates northeastward past St. Lawrence Island and into Norton Sound. Over most of the shelf region the tide is mainly semidiurnal, but in Norton Sound the diurnal tides predominate. A number of amphidromic systems are formed on the eastern Bering shelf and in Norton Sound as a result of the interference of tides propagating from various directions. Various authors have different opinions about the details of the tidal height and current distributions, including the exist-

ence, shape, and location of amphidromic systems. Some of the controversy can be resolved with the recently acquired data; the rest will require more observations and/or complete models.

NUMERICAL MODELS

Several numerical models have been applied to the tides in the Bering Sea. They are of two basic types: vertically integrated models which simulate horizontal distributions and three-dimensional models which simulate vertical variations as well. Some of these models are still under development.

The vertically integrated models by Hastings (1976) and Sünderman (1977) superimpose uniformly spaced grids over the Bering Sea Shelf and the entire Bering Sea, respectively. Finite difference approximations to the dynamic equations with quadratic bottom friction are solved over the grids as initial value problems in time with lateral friction included to stabilize the calculation. The tides enter the models as boundary conditions along the open boundaries; time series of sea level at the open boundaries drive the motion in the interior with zero flux and no-slip conditions imposed along coasts. The imposed time series are derived from harmonic constants interpolated from observed values at islands and on coasts. After integrating the models through an initial transient, the motion can be analyzed for tides and tidal currents. Since Hastings (1976) does not take this step, we show only results of the model by Sünderman (1977), which is limited to the M_2 semidiurnal constituent (Fig. 8-2) and its consequences in time-averaged properties and higher harmonics.

A distinctly different, vertically integrated model is under development by Preisendorfer (1979), who uses Bristol Bay to illustrate a new technique. The technique involves computing the linear response of a region to a long wave of a given frequency, such as a tidal constituent, as a synthesis of responses to simple subregions. Since realistic boundary conditions were not used in the Bristol Bay example, the results of the calculation may be considered preliminary. Further work on the technique is planned (Preisendorfer, personal communication).

A three-dimensional model has been developed by Leendertse and Liu (1977) and Liu and Leendertse (1978, 1979) to predict tides and wind-driven currents on the southeast Bering shelf for the prediction of oil spill trajectories and for risk analysis. The grid for the model (Fig. 8-3a) is uniform in the horizontal dimensions but packed in the vertical dimension to allow higher resolution of the pycno-

Figure 8-2. Charts of (a) coamplitude (cm), (b) cophase (Greenwich lag in degrees), and (c) tidal current ellipses (cm/sec; radial line within each ellipse represents the tidal velocity at Greenwich transit) from the vertically integrated model by Sünderman (1977) of the M_2 tide in the Bering Sea. For the finite-difference model: the grid size is 75 km; the time step is 223.5 sec; the bottom drag coefficient is 0.003; and the lateral viscosity is 10^9 cm²/sec. The numbers appearing in boxes are observed values. (Reproduced with permission from J. Sünderman, 1977, *Deutsche Hydrographische Zeitschrift*, 91-101, Figs. 3, 4, 5).

cline. The dynamic variables are the horizontal and vertical velocity components, temperature, salinity, density, pressure, and the energy density at sub-grid scales; a passive contaminant can also be included in the model. The large-scale (resolved by the grid) variables satisfy a relatively complete set of nonlinear dynamic equations averaged over each vertical layer and specialized to the extent that the hydrostatic equation is used in the vertical direction. The subgrid scale turbulence satisfies a one-equation closure model in the turbulent energy density.

Within the dynamic equations are viscous and diffusive terms. The associated viscosities and diffusivities in the horizontal direction have contributions from the large-scale motions through the local vorticity gradient. The contributions of the subgrid turbulence for each direction depend upon a turbulent Richardson number which includes the local density gradient and the local subgrid energy density. The formulas for the turbulent contributions are modified from Mamayev (1958) and include empirical coefficients chosen to produce reasonable behavior of the turbulence in the presence of stratification. Liu (personal communication) indicates that newer versions of the model under development will use a different formulation for the turbulent viscosities and diffusivities which eliminates many of the empirical coefficients.

The motion in the lowest layer is subject to bottom boundary conditions. The bottom is assumed to be impervious and insulating so that zero flux of heat, salt, and water occurs through the bottom. The bottom boundary condition for momentum is a quadratic friction law in which the Chezy coefficient is adjusted to match model currents with observed currents. At the free surface, a uniform wind stress can be imposed over the region to produce wind-driven currents in addition to the tidal motion. The model is integrated as an initial value problem with predicted tide-level time series specified along the open boundaries; these time series are obtained through interpolation of harmonic constants derived from bottom pressure data.

To simulate tides and tidal currents on the south-east Bering shelf, Liu and Leendertse (1978, 1979) chose a period, 16-18 June 1976, for which extensive observations existed, some collected specifically in support of this modelling effort. The model was run for a total of 63 hours, which included an initial transient preceding the comparison period. Time series of sea level were then Fourier analyzed for composite semidaily and daily tides from which cotidal charts (Fig. 8-4a,b) were constructed. A more

Figure 8-3a, b. (a) The horizontal grid and (b) a section showing the vertical grid of the three-dimensional, turbulent energy model of the south-eastern Bering Shelf by Liu and Leendertse (1978, 1979). The horizontal grid spacing is 21.82 km; the vertical spacing in this 14-layer model is variable to allow high resolution in the pycnocline. The time step for the model is 180 sec, and the Chezy coefficient of bottom drag is 700 ($\text{cm}^{1/2}/\text{sec}$). (Reproduced with permission from S. K. Liu and J. J. Leendertse 1979).

Figure 8-4a, b. Theoretical cotidal charts (amplitudes in cm, phases in degrees relative to the start of the simulation) for (a) the composite diurnal tide and (b) the composite semidiurnal tide obtained from the three-dimensional, turbulent energy model by Liu and Leendertse (1978) of the southeastern Bering shelf. (Reproduced with permission from S K. Liu and J. J. Leendertse 1979.)

detailed analysis into individual tidal constituents was not possible because of the short series length. After model coefficients were adjusted to match current data, a set of time series (Fig. 8-5a-c) of currents and water properties was produced.

A similar three-dimensional model is under development for Norton Sound (Liu, personal communication). It includes improved equations for the subgrid-scale turbulence. The model can also impose sea ice on the water; ice in the north Bering Shelf region can be expected to modify the tidal and wind-driven responses significantly.

OBSERVATIONS

Offshore pressure and current observations, made on the east Bering Shelf (Fig. 8-6) during 1975-78, are a major new source of information about tidal heights and currents. These data have been analyzed¹ for harmonic constants and tidal current ellipses; the analysis will be presented in this section. To these results have been added historical harmonic constants from the International Hydrographic Bureau (1966) and results of analysis of previously unpublished data from the National Ocean Survey and the U.S. Geological Survey.

Tidal observations

Empirical cotidal charts (Figs. 8-7 through 8-10) have been constructed for the four largest tidal constituents: M_2 , N_2 , K_1 , and O_1 . These charts were constructed as follows: Cotidal lines were drawn by interpolation where the station spacing is relatively small compared with the spatial scales of a given tidal constituent. Elsewhere, the positions of cotidal lines were estimated either by shifting the distribution of lines from numerical models to match observed values, or by estimation with reference to hydrodynamic considerations. Like all empirical cotidal charts, these charts are somewhat subjective. On each of the cotidal charts, regions of particular uncertainty are denoted by question marks.

An example of the procedure for drawing charts can be seen from the portion of the M_2 chart (Fig. 8-7) around St. Lawrence Island. Along the northern shore of the island, M_2 harmonic constants can be found for enough stations to document the westward progression of phase without reference to models. Along the southern shore of the island there are fewer

stations, and the convergence of M_2 cophase lines at Southeast Cape was drawn with reference (Fig. 8-2) to the M_2 model by Sündermann (1977). The records along the northern coast of St. Lawrence Island provide harmonic constants for M_2 derived from a high water/low water analysis (see Appendix). These data are not sufficient to allow a determination of harmonic constants of other constituents; the cotidal charts of those constituents in this region are therefore more speculative than that of M_2 .

The cotidal charts show the complex distribution of tidal amplitudes and phases over the eastern shelf region. Largest amplitudes for all constituents are found along the Alaska Peninsula, interior Bristol Bay, and in Kuskokwim Bay. The smallest tides are found on the outer shelf and north of St. Lawrence Island. A number of amphidromes are present; there is a semidiurnal amphidrome in Norton Sound and a virtual amphidrome at Cape Newenham, while diurnal amphidromes are found on the southeast shelf between Nunivak Island and the Pribilof Islands and west of Norton Sound. The last is evident only for the K_1 constituent. The situation is more ambiguous for the O_1 tide north of St. Lawrence Island, possibly because the longer wavelength of O_1 would place an amphidrome farther west of the offshore stations. The phase difference between K_1 and O_1 changes rapidly at stations in this area, indicating that the structures of the two constituents are significantly different.

The S_2 tide, normally the second largest semidiurnal constituent, is unusually small throughout the Bering Sea. Typically, S_2 amplitudes range from 1 to 3 cm. The cotidal charts by Bogdanov (1961) and Bogdanov et. al. (1964) for the North Pacific Ocean show an S_2 amphidrome just south of the Aleutian Islands. Perhaps the small S_2 amplitude where tides propagate into the Bering Sea produces small S_2 amplitudes throughout the region. Since analysis of data from the Bering Sea did not produce stable S_2 harmonic constants, a cotidal chart for S_2 could not be constructed. A result of the small S_2 is that the fortnightly inequality, or semidiurnal spring-neap cycle, is much less important than the parallactic inequality due to the variation in the moon's distance from the earth. This cycle is equal to the period of the moon's orbit, 27.55 days, and is manifested in the beat of N_2 against M_2 .

Tidal type may be classified (e.g., Defant 1961) by the value of the ratios of the sums of amplitudes of principal diurnal constituents K_1 and O_1 the principal semidiurnal constituents M_2 and S_2 .

¹ Details of the analysis procedure and results are given in the Appendix.

Figure 8-6. Location of stations used in construction of cotidal and tidal current ellipse charts. Historic stations with names had harmonic constants available; those without names were used to estimate M_2 from high and low water analysis.

Since S_2 is so small in the Bering Sea we have substituted N_2 in the equation:

$$F = \frac{K_1 + O_1}{M_2 + N_2}$$

Values of F less than 0.25 denote regular semidiurnal tides, with two high and two low waters per day of approximately the same height. The mixed, predominantly semidiurnal type has values of 0.25-1.5 and is characterized by two high and two low waters per day, but with large diurnal inequalities in heights. Mixed predominantly diurnal tides, $F = 1.5-3.0$, have only one high and one low water per day when the

moon is near its maximum declination. Regular diurnal tides have values of F greater than 3.0 and usually have only one high and one low water per day.

In most areas of the eastern Bering Sea the tide is the mixed semidiurnal type. The regular semidiurnal type is found in the Bering Strait and in a small area near the diurnal amphidrome south of Nunivak Island. The mixed diurnal type is found along the Aleutian chain, probably near Cape Newenham and in eastern Norton Sound. Regular diurnal tides are found in the vicinity of the semidiurnal amphidrome in Norton Sound.

Figure 8-7. M_2 cotidal chart based on the new observations. Dots represent locations of stations used in construction of the chart. Solid lines are cophase lines referred to Greenwich. Dashed lines are coamplitude, in centimeters. Areas of major uncertainty are denoted by question marks.

Figure 8-8. Same as 8-7 but for N_2 .

Figure 8-9. Same as 8-7 but for K_1 .

Figure 8-10. Same as 8-7 but for O_1 .

Figure 8-11. Predicted tides for the month of April 1980 for selected stations, illustrating tidal types found in the Bering Sea. Predictions used M_2 , N_2 , S_2 , K_1 , O_1 , and P_1 .

Examples of different tide types are shown in Fig. 8-11, which shows predicted tide curves for the month of April 1980. Station BC9, southwest of Nunivak Island near the diurnal amphidrome, has an F -value of 0.25 and is mainly semidiurnal. The lunar perigee is on the 14th with the largest range on about April 18 and the smallest at the beginning and end of the month. BC2 in Bristol Bay has the mixed semidiurnal type with $F = 0.79$. The inequalities are largest after maximum declination of the moon on the 7th and 20th. BC20, near St. Matthew Island, also has a mixed semidiurnal tide, $F = 1.02$. Note that while at BC2 the diurnal inequality is mainly in the low tide, at BC20 it is in both the high and low waters. This is caused by differences in the phase relationships between the major diurnal ($O_1 + K_1$) tides and M_2 . Finally, the regular diurnal tide type is found at LD5, near the semidiurnal amphidrome in

Norton Sound. Here $F = 16.3$ and the fortnightly tropic-equatorial tide cycle, caused by the declination of the moon, is evident.

Tidal current observations

Harmonic constants of current-meter data are subject to more variability in time and space than those of pressure-gauge data, which are generally quite stable. A variety of factors affect observed tidal current velocities, including local bathymetry and depth, shear in the water column, and increased damping and friction in the upper layers due to ice cover. In addition, the Savonius rotor current meters used in this study were occasionally subject to errors due to wave-induced mooring motion, which resulted in higher recorded speeds, and biological fouling of the rotor by algae and various types of drifting

marine organisms, resulting in lower recorded speeds.

To examine temporal variability, successive 29-day harmonic tide analyses were performed on all available current-meter data. Analyses were performed on the east and north components of velocity. Several moorings had decreases in tidal amplitudes corresponding to times of ice cover. Generally, the semidiurnal constituents were reduced more than the diurnal constituents, and reductions were greatest for the upper meters. Phases were also affected. Fig. 8-12 shows variations in the M_2 constituents of the east component of current at station NC24 at depths of 24 m and 40 m for a one-year period beginning 19 September 1977 (the north component exhibited similar behavior). Ice began forming in the area of the mooring in late November and early December of 1977 (Fleet Weather Facility 1977 and 1978); this ice formation was associated with a decrease in the M_2 amplitude at the upper meter to 18 cm/sec. However, the amount of ice over the mooring and the location of the ice edge varied until February 1978, when the ice edge moved much further south. Amplitudes were lowest in February and March during the time of extensive ice cover and then increased again as the ice broke up. Amplitudes were reduced by almost 40 percent, from 22 cm/sec in the early fall to about 14 cm/sec during late winter. During the summer 1978 amplitudes returned to over 20 cm/sec. The effect was much less at the lower meter. Amplitudes were decreased from 15.4 cm/sec to about 12.5 cm/sec or less than 20 percent.

Figure 8-12. Variations in the M_2 constituent of the east component of current at station NC24 for a one-year period beginning September 1977. Upper meter was at a depth of 24 m (19 m for the last two months). Lower meter depth was 40 m.

Phases at both meters also varied. At the upper meter phase decreased from 204° to 190° while at the lower meter phase increased from 201° to about 217° . This was associated with a shift in the ellipse axis orientation toward the left at the lower meter while the upper-meter ellipse remained steady. The K_1 amplitudes were also reduced, although to a lesser extent than M_2 . It is reasonably certain that these changes are associated with ice cover. Whether these phenomena were due more to a reduction of mooring noise resulting from wave attenuation by the ice or to frictional effects of ice cover remains unknown. Presumably, mooring noise would increase the amplitudes of all frequencies by more or less the same amount; moreover, phases would not be affected. Nevertheless, mooring noise is probably an important consideration, especially during the stormy fall and spring months.

The component harmonic constants for M_2 , N_2 , K_1 , and O_1 were converted to an ellipse representation to give amplitude H and phase G for the major axis, direction of the major axis, amplitude of the minor axis, and sense of rotation. Usually the analysis for the first 29 days of the mooring period was used, to minimize effects of fouling on the current harmonic constants. Table 8-1 gives the results of the analyses for a representative distribution of stations. Some stations were not included because of close proximity to one or more of those listed.

Plots of ellipses for stations listed are shown in Figs. 8-13 and 8-14 for the M_2 and K_1 constituents, respectively. Ellipses are centered at the station locations with lines from the center representing the constituent current vector for the time at which the equilibrium constituent passes the Greenwich meridian.

The M_2 ellipses (Fig. 8-13) exhibit clockwise rotation at most stations south of St. Lawrence Island. The exception is BC14 near the Alaska Peninsula, which has nearly rectilinear motion. Axes are generally aligned in the direction of wave propagation, although topographic constraints can modify this, as at BC18, just south of Nunivak Island. Away from the influences of land, ellipses are more nearly circular. Amplitudes are typically 15-30 cm/sec on the open shelf, although somewhat higher speeds would be expected at the surface. In Norton Sound and near St. Lawrence Island, rotation is anticlockwise. M_2 current speeds are small in Norton Sound.

The K_1 ellipses (Fig. 8-14) are narrower than M_2 . The orientation is aligned with flow in and out of the major embayments, Bristol Bay and Norton Sound. Over the outer shelf, the rotation is clockwise but becomes generally anticlockwise in the mid-shelf

Figure 8-13. M_2 current ellipses for stations listed in Table 8-2. For stations with records from two depths the ellipse for the deeper meter is plotted. Ellipses are centered on station location; line from center indicates constituent current vector when the M_2 Greenwich equilibrium phase angle is 0° . Arrows indicate sense of rotation.

Figure 8-14. Same as 8-13 but for K_1 .

TABLE 8-1

Results of 29-Day harmonic analyses for selected current-meter records, in ellipse representation¹

Station	Depth	Meter	Depth	Lat N	Long W	O ₁			K ₁			N ₂			M ₂													
						major			min			major			min			major			min							
						Yr	JD	H	G	D	H	R	H	G	D	H	R	H	G	D	H	R						
BC3	114	20	55	02	165	10	76	150	7.9	245	52	1.7	C	9.8	279	57	1.0	C	6.5	26	37	1.8	C	26.4	71	38	7.6	C
BC3	114	100	55	02	165	10	76	150	7.8	265	52	1.2	C	10.4	281	48	2.1	C	4.9	11	50	0.6	C	21.0	61	47	4.4	C
BC13B	115	100	55	30	165	49	76	158	5.6	272	71	2.2	C	8.0	289	68	2.8	C	5.3	3	50	2.2	C	18.9	67	54	9.6	C
BC17	104	96	56	34	167	34	76	266	6.5	312	109	1.8	C	9.1	326	105	4.3	C	6.4	27	61	4.0	C	15.9	81	53	10.5	C
BC4	51	30	58	37	168	14	75	251	7.4	330	140	3.1	C	11.7	350	133	6.2	C	9.6	45	49	7.2	C	27.3	118	50	22.2	C
BC4	51	47	58	37	168	14	75	251	7.3	331	128	2.6	C	10.9	0	131	4.8	C	4.9	28	46	3.3	C	20.2	107	51	15.1	C
BC9	41	33	59	12	167	43	77	133	8.5	332	130	0.9	C	13.0	359	134	2.8	C	7.0	60	47	5.1	C	18.7	120	51	14.6	C
BC18	31	20	59	40	167	07	77	132	12.7	333	113	0.6	A	20.5	356	116	0.1	A	6.3	130	109	4.8	C	21.2	186	103	13.8	C
BC16	50	37	57	59	165	16	77	123	7.7	308	102	1.3	A	11.4	328	108	1.5	A	6.8	69	69	5.8	C	20.8	101	52	15.1	C
BC2	65	50	57	04	163	22	75	310	8.6	292	77	1.1	A	14.1	309	81	1.6	A	5.9	76	81	2.9	C	20.4	145	80	12.1	C
BC14	51	20	56	03	161	50	76	152	13.5	292	68	0.5	A	21.7	312	70	0.5	A	12.2	98	69	0.1	A	35.4	159	67	0.7	C
BC20	64	51	60	26	171	05	78	202	4.5	186	340	2.4	C	6.7	214	353	4.5	C	4.5	68	68	2.7	C	13.7	136	61	9.5	C
BC21	42	20	60	24	169	10	78	202	8.0	188	352	2.4	C	12.5	219	0	4.6	C	8.0	86	68	3.0	C	28.6	150	63	17.1	C
BC21	42	32	60	24	169	10	78	202	7.9	187	351	2.9	C	11.6	221	353	4.1	C	6.9	87	61	2.5	C	21.1	139	64	9.6	C
NC24	48	19	61	48	170	26	78	203	3.4	212	358	2.7	C	5.5	263	35	4.4	C	6.4	144	86	3.7	C	21.1	206	92	12.5	C
NC24	48	39	61	48	170	26	78	203	2.9	189	335	2.7	C	4.6	274	33	3.9	C	4.1	120	103	1.3	C	14.1	191	103	6.1	C
LD1	14	10	62	30	166	07	78	204	10.9	261	19	0.2	C	19.8	290	34	1.2	A	5.9	268	28	3.3	A	17.5	56	298	12.8	A
NC17	26	17	62	53	167	05	78	165	8.8	271	39	0.1	A	12.8	310	41	1.7	A	4.6	348	305	1.0	A	12.1	48	301	2.3	A
LD2	28	24	63	13	168	35	78	203	4.5	220	20	0.3	C	7.5	29	11	0.7	C	4.5	12	201	1.7	A	15.4	310	211	8.5	A
LD3	37	33	64	00	168	00	78	204	1.9	342	63	0.1	C	4.0	24	73	1.7	C	1.3	39	301	1.1	A	4.9	78	322	3.5	A
LD4	20	16	64	47	166	50	78	205	3.9	144	294	1.6	C	7.5	222	309	0.9	C	2.7	358	17	0.4	C	6.8	74	356	4.0	A
NC20	19	5	64	00	165	29	77	189	18.5	315	70	3.0	C	33.0	22	70	5.7	C	2.0	244	126	0.5	A	7.5	303	155	0.9	A
NC14	32	21	64	22	165	22	76	234	14.5	327	93	0.4	C	27.7	27	94	0.9	C	1.9	167	90	0.5	C	3.3	182	85	0.5	C
LD5	27	20	64	08	163	00	78	206	9.3	335	66	1.7	A	22.7	25	67	4.8	A	3.0	167	0	2.4	A	8.0	244	331	6.7	A

¹ Amplitudes H are cm/sec, phases G are referred to Greenwich, and direction D of major axis is compass degrees. C refers to clockwise rotation, A to anticlockwise. To obtain phase and direction of minor axis, add 90° to major axis phase if rotation is clockwise, or subtract 90° if anticlockwise.

region of Bristol Bay (stations BC16, BC2, BC14) and in the approaches to Norton Sound (stations LD1, LD2, NC17). Typical amplitudes over the southeast shelf are 10-20 cm/sec. Largest K_1 currents are found within Norton Sound where amplitudes range from 20 to over 30 cm/sec. West of Norton Sound amplitudes are very small.

Ellipses for the smaller N_2 and O_1 constituents have not been plotted. Generally they are similar to the M_2 and K_1 respectively. O_1 major axis amplitudes are 60-75 percent of K_1 in the southeastern Bering Sea and 40-60 percent in the Norton Sound region. N_2 is about 25-40 percent of M_2 throughout the Bering Sea.

DISCUSSION

Both models and observations show that the largest tides in the Bering Sea occur on the inner southeast Bering shelf. As shown in the figures, there is general agreement that the tides propagate as free waves onto the southeast shelf from the deep basin. In the vicinity of the Alaska Peninsula, the exponential decay of amplitude northward from shore and a comparison of the tidal heights and currents at BC14 indicate that the semidiurnal and diurnal tides are Kelvin² waves near the peninsula, the semidiurnal wave is progressive as at BC2, where the M_2 phase difference is 12° , and at BC14, where the current is nearly in phase with the tide. Conversely, the K_1 tide has a much larger standing wave component, with a phase difference of 64° at BC2 and about 40° at BC14. When the Kelvin waves propagate into the inner shelf, there is an interesting difference between the reflected waves for the two tidal species. The critical latitudes of the diurnal tides ($\sim 30^\circ\text{N}$) lie south of the Bering Sea; diurnal tides can therefore obey only Kelvin-wave dynamics on the open shelf. The reflection of each diurnal tidal constituent produces a classic amphidromic system on the southeast shelf with the amphidromic point well offshore.

The critical latitudes of the semidiurnal tides (75°N for M_2) lie north of the Bering Sea. As a

² Kelvin and Sverdrup waves are long waves under the influence of the earth's rotation. In the northern hemisphere Kelvin waves propagate with the coast on the right and amplitude decreasing exponentially away from the coast. Current motion is rectilinear (in the absence of friction). Sverdrup waves (also called Poincaré or inertial-gravity waves) have horizontal wave crests and the currents form a clockwise-rotating ellipse. Sverdrup waves can exist only where the wave period is less than the inertial period. In areas where both Kelvin and Sverdrup waves can exist, the actual wave is some combination of the two. For further discussion, the reader is referred to a text such as *The Oceans* (Sverdrup et al. 1942).

result, semidiurnal tides can obey Sverdrup-wave as well as Kelvin-wave dynamics. The presence of a virtual amphidrome near Cape Newenham for each semidiurnal constituent suggests that a major portion (if not all) of the incident, semidiurnal Kelvin waves are dissipated or converted on reflection into Sverdrup waves. Perhaps the acute apex angle of Kvichak Bay, together with the presence of sharply protruding peninsulas, may produce an efficient conversion to semidiurnal Sverdrup waves. A large part of the semidiurnal tidal energy may also be dissipated over the extensive mud flats and shoals of the Kvichak and Nushagak Bays and Rivers. On the open shelf, the broad semidiurnal tidal ellipses have approximately $45\text{--}50^\circ$ phases relative to the tidal heights in the direction of propagation. This fact and the fact that the offshore cophase lines are roughly parallel to the coast and widely separated indicate that the semidiurnal tides act as a standing Sverdrup (Poincaré) wave in this region due to cooscillation in Kuskokwim Bay. Further cooscillation is evident at BC11, just southwest of Nunivak Island, where the M_2 tide actually leads that at stations farther seaward.

The *Coast Pilot*, published by the U.S. Department of Commerce (1964), notes that to the north of the southeast Bering shelf the currents in Etolin Strait between Nunivak Island and the mainland are sufficiently strong to prevent ice formation in winter. This may be due to tidal currents associated with the great differences in tidal phases between the north and south sides of Nunivak Island.

There is much less consensus about the tides in the north Bering Sea. The tides around St. Lawrence Island are particularly confusing. Harris (1904) shows the tide progressing from west to east along the south shore and then east to west along the north shore, so that all phases of the tide are found around the island. Leonov (1960) has branches of the tide progressing through the passes on the west and east sides of the island and meeting on the north side together with the tide from the Arctic Ocean; he discusses areas of convergence and divergence and abrupt changes in velocity as a result of the meeting of the tides of the Pacific and Arctic Oceans. Conversely, the Office of Climatology and Oceanographic Analysis Division (1961) chart shows very little effect on the tide from St. Lawrence Island with a fairly smooth progression of the tide northward across the shelf and through the Bering Strait. Coachman et al. (1975) inferred the presence of an amphidrome south of St. Lawrence Island from tidal differences of stations listed in the tide tables.

According to Sünderman's (1977) charts (Fig.

8-2), cophase lines converge on Southeast Cape, and there is probably cooscillation in the bight on the south side of the island with a reversal of phase progression. This may be responsible for the large phase differences noted by Coachman et al. (1975). The tide appears to progress through the pass between Northwest Cape and Siberia and thence west to east along the north shore of St. Lawrence Island to the vicinity of Northeast Cape, where it meets the wave progressing north along the Alaskan mainland. This results in nearly 180° phase difference between the M_2 currents at LD1, near the Yukon Delta, and at LD2 near Northeast Cape. That is, current is northerly at LD1 when it is southerly at LD2. Station NC17, between these two stations, is in a transition zone; most motion is cross-channel.

Coachman et al. (1975) discussed measurements made at a current-meter mooring 90 km south of the Bering Strait. Tidal currents were mainly semidiurnal with amplitudes of 1-12 cm/sec and the current ellipse orientation was generally northeast-southwest to north-south. In the Bering Strait, Ratmanoff (1937) and Fleming and Heggarty (1966) did not find noticeable tidal currents while Bloom (1964) and Coachman and Aagaard (1966) found that tidal currents modulated the net northward flow.

Semidiurnal currents are especially small in Norton Sound south of Nome, for example at stations NC14 and NC20. These stations are in an antinode one-half wavelength from the head of the bay. Generally, semidiurnal tides and tidal currents are small throughout the Norton Sound and Bering Strait region, perhaps as a result of frictional dissipation across the broad Bering Sea shelf and Kelvin-wave dynamics (i.e., the Norton Sound amphidrome and small amplitudes to the west of a northward-progressing wave).

The diurnal tides appear to be simpler than the semidiurnal tides because they are restricted to Kelvin-wave dynamics and have longer wavelengths. Tidal height/tidal current phase relationships indicate predominantly standing-wave characteristics in southeast Bering Sea and Norton Sound although between St. Lawrence Island and the Alaskan mainland the tide is more progressive. Thus the diurnal tide in the deep western Bering Sea basin cooscillates with Bristol Bay and Norton Sound. Currents are highest in Norton Sound because of the shallower depth, even though the diurnal tides are higher at the head of Bristol Bay. Between St. Lawrence Island and the Bering Strait, the diurnal tides virtually disappear, again due to Kelvin-wave dynamics and dissipation.

While there is good agreement between the models and the observations on the major features of the

tides of the eastern Bering Sea shelf, there are some differences. For instance, the M_2 cotidal chart of Sünderman (1977) has systematically higher amplitudes and later phases than the observations on the southeast shelf, but lower amplitudes in Kvichak Bay. Also, the actual amphidrome in Norton Sound appears to be further ashore. These differences may be due to the coarse grid (75 km) and high horizontal eddy viscosity (10^9 cm²/sec) used in the model.

It is more difficult to evaluate the Liu and Leendertse (1978, 1979) model (Figs. 8-3-5) since many of the observations were used for boundary conditions and for tuning empirical parameters. The model and observations are therefore not independent. The cotidal charts (Fig. 8-4) present composite tides based on a Fourier analysis of 50 hours of computed data; therefore the cotidal charts represent sums of constituents within each species, with an arbitrary composite phase rather than one referred to Greenwich. For these reasons, no attempt will be made here to assess the accuracy of this model.

APPENDIX

Moorings usually consisted of one or two Aanderaa RCM-4 current meters on a taut wire, with the upper current meter at a depth of about 20 m, just below the subsurface float; the deeper one was below the pycnocline, about 10 m above the bottom. In the Norton Sound area, where depths are generally less than 30 m, moorings contained only one meter. The Aanderaa TG2 or TG3 pressure gauge, if present, either was placed in a well within the anchor or was attached to acoustic release, just above the anchor. Pressure-gauge resolution is claimed by the manufacturer to be better than 0.5 cm. Individual moorings were in place for periods of up to one year, with observations of some locations spanning as much as three years.

Data were processed by methods similar to those described in Charnell and Krancus (1976). No correction was made for atmospheric pressure in the pressure-gage data. In Aanderaa current meters, the recorded speed is an average over the data interval (15, 20, 30, 40, or 60 minutes) while direction is essentially instantaneous. To remove the phase difference between speed and direction, speeds at times T_n and T_{n+1} were averaged to give the speed corresponding to the direction at T_n before converting to east and north components of velocity. The data were low-pass filtered to remove noise (i.e., energy at frequencies >0.5 cycle/hr) and resampled at hourly intervals. A second-order polynomial was then used to interpolate to even hours.

The Munk-Cartwright response method (Munk and Cartwright 1966) was used for tidal height analysis using procedures based on those suggested by Cartwright et al. (1969). The pressure record from Station BC20 was selected as a reference for all other stations because of its length (300 days) and location (near the center of the study area), and because the tide there is thought to be representative of that entering the shelf from the deep basin of the Bering Sea. The tide potential was used as a reference for the BC20 pressure record, using weights of 0, ± 2 , and ± 4 days for the diurnal and semidiurnal bands. For the other stations the reference series was a complex prediction based on the sixteen largest diurnal and semidiurnal harmonic constituents derived from the BC20 analysis using weights of 0, ± 2 days.

Results of high and low water analyses for many locations in the Bering Sea were obtained from the National Ocean Survey. High and low water analysis

gives the mean tide range and Greenwich high and low water lunitidal intervals HWI and LWI. According to Schureman (1958), if the tide is semi-diurnal, the M_2 amplitude can be estimated by multiplying the mean range by 0.47. The phase may be found by:

$$M_2^\circ = \frac{1}{2}(\text{HWI} + \text{LWI}) \times 28.984 + 90^\circ.$$

For current-meter data, a 29-day harmonic analysis based on Schureman (1958) was used. Constituents O_1 , K_1 , N_2 , M_2 , and S_2 are derived directly, and other constituents are inferred from these on the basis of equilibrium relationships. The harmonic method was used for currents because uncertainties in data quality and seasonal variations obviated the use of the slightly more accurate response method.

The results of these analyses are given in Table 8-1 (for current-meter data, the four major constituents in an ellipse representation) and in Table 8-2

TABLE 8-2
Results of response analyses for the new pressure-gage observations¹

Station	Lat, N		Long, W	O_1		P_1		K_1		N_2		M_2		S_2		Start Date Yr JD		Length Days
				H	G	H	G	H	G	H	G	H	G	H	G			
BC20	60	26	171	05	9.8	310	5.8	323	18.1	326	6.9	121	20.5	171	2.2	249	77 260	300
BC3	55	01	165	10	26.4	304	13.3	317	40.9	319	15.8	40	41.9	89	3.2	2	76 077	73
BC13B	55	30	165	49	23.8	311	11.4	322	34.4	325	13.2	52	35.5	106	1.8	327	76 158	114
BC13D	55	47	165	23	23.0	312	11.0	324	33.4	327	15.0	56	39.0	109	1.4	314	77 252	131
BC10	57	17	169	33	17.3	319	8.2	330	24.9	333	8.7	77	24.9	131	1.5	266	76 153	101
BC4	58	37	168	14	6.3	305	3.9	300	12.4	303	11.1	98	33.4	151	1.8	252	75 250	58
FX2	58	32	167	56	4.0	286	1.8	284	8.9	288	11.8	104	33.8	158	1.9	251	78 200	64
BC9	59	13	167	42	2.4	220	3.1	253	9.7	258	12.4	108	36.7	164	2.0	246	76 269	230
BC11	59	42	167	15	10.6	165	6.0	204	18.3	207	11.9	98	35.9	155	2.8	224	76 154	102
BC21	60	23	169	11	7.4	271	5.3	294	16.6	297	10.1	136	30.9	189	2.9	265	77 260	246
BC7	55	42	163	01	31.3	318	16.0	332	49.0	335	24.5	81	71.4	134	1.4	45	76 080	70
BC2	57	04	163	22	19.0	358	9.3	11	28.3	13	14.7	102	45.2	157	0.5	343	76 151	200
BC15	57	39	162	42	21.7	25	9.9	45	29.9	48	11.1	116	36.2	168	1.3	257	77 257	129
LD1	62	30	166	07	17.9	286	10.3	319	31.7	324	13.2	274	46.1	328	6.8	49	78 204	54
NC17	62	53	167	05	10.7	303	6.2	336	19.0	341	7.4	274	25.6	330	3.0	121	77 263	293
NC18	63	09	168	23	4.4	339	2.9	351	8.9	356	4.7	272	22.4	324	2.6	59	76 245	120
LD2	63	13	168	35	4.6	338	2.6	356	8.1	359	6.2	261	26.6	319	5.3	58	78 203	54
LD4	64	47	166	50	2.3	75	1.1	218	4.2	222	2.3	47	4.9	138	0.6	324	78 205	55
NC10	65	45	168	27	0.5	228	0.6	277	2.7	296	2.5	147	12.0	213	3.5	284	77 202	26
GEO	64	00	165	30	9.5	23	4.8	66	14.0	71	4.4	349	13.0	44	2.5	131	77 189	60
PROBE																		
LD5	64	08	163	00	16.9	60	10.4	106	32.1	110	1.0	186	2.0	233	0.2	277	78 206	41
UN-ALAKLEET	65	53	160	47	24.0	55	13.9	102	42.9	108	5.2	164	17.3	222	3.0	326	77 220	95

¹ Amplitude H in mbar of pressure. 1 mbar equals $1.007 \pm .003$ cm of sea water in the Bering Sea. Phases G are referred to Greenwich.

(for pressure gauges, the six major constituents). Harmonic constant amplitudes H are cm/sec for currents and mbar for pressure-gauge data. For this region, 1 mbar equals $1.007 \pm .003$ cm of sea water. Phases are referred to Greenwich.

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Section II

Ice Distribution and Dynamics

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Recent Fluctuations in Sea Ice Distribution in the Eastern Bering Sea

9

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ABSTRACT

The eastern Bering Sea shelf (~1000 km long $\times$ 500 km wide) is ice covered in winter but ice free in summer. Time series of weekly percent ice coverage are presented, illustrating details of this phenomenon for the period 1973-79. Advances and retreats of the ice edge are correlated with fluctuations in sea and air temperatures, with surface winds, and with regional meteorological events. The period 1973-79 is shown to be a time of extreme fluctuations with 1973-76 characterized by below-normal temperatures and above-normal ice cover under northerly winds, while 1976-79 was a period of strong rise in temperatures and retreat of the ice pack under winds shifting to southerly.

INTRODUCTION

The southeast Bering Sea shelf is a relatively shallow (shelf break ~150 m) but wide (~500 km) region (Fig. 9-1) that is seasonally covered with ice. During a typical winter, the ice advances about 1,000 km south from the Bering Strait to the shelf break (Fig. 9-1). This advance occurs primarily by freezing of seawater within the Bering Sea (Leonov 1960) and does not imply advective advance through the Bering Strait (Tabata 1974). In spring, about 63 percent of the ice melts within the basin (Lisitsyn 1960), and the remainder leaves by way of the various passes and straits.

In addition to the strong seasonal cycle in ice coverage, large multiyear variations have been observed in the Bering Sea. Walsh and Johnson (1979), who considered the extent of sea ice in the Northern Hemisphere for the years 1953-77, show year-to-year ice fluctuations exceeding 5° latitude (~340 km) in nearly all the seas surrounding the Arctic Ocean. However, they show that the standard deviation of the departure from the monthly

mean extent of sea ice is greater in the Bering Sea along 170°W (Fig. 9-1) than in the other high latitude seas.

As an example of one of these fluctuations, Kukla and Kukla (1974) showed that the onset of the decline of sea-surface temperatures (SST) in the early 1970's on the Bering Sea shelf coincided with anomalous southward penetration of the ice pack. McLain and Favorite (1976) related the fall in SST during this period to changes in the atmospheric circulation which caused northerly winds over the Bering Sea. Niebauer (1978) has related a subsequent rise in SST in the late 1970's to, again, changes in the atmospheric circulation which caused southerly air flow over the region. This chapter analyzes the remarkable decrease in iciness as related to this recent short-term climatic fluctuation and rise in SST reported by Niebauer (1980; Chapter 3, this volume).

On similar time scales, Sater et al. (1974) and Rodgers (1978) have shown that ice conditions in the Beaufort Sea are correlated with meteorological conditions such as sea-level pressure and wind direction. Rodgers has related light ice summers to more frequent southerly surface winds; the winds are reversed in heavy ice summers. For a recent bibliography of authors who have described interannual sea-ice fluctuations of several hundred kilometers in other high latitude seas surrounding the Arctic Ocean, see Walsh and Johnson (1979).

The following sections present a description and analysis of the seasonal ice cycle derived from weekly mean percent ice cover for the eastern Bering Sea for 1973-79. The multiyear variations are then discussed in relation to short-term climatic variations and to fluctuations in SST, air temperatures, and surface winds.

Figure 9-1. Chart of the Bering Sea showing the bathymetry of the eastern shelf. Examples of the southern ice limit for April 1976 and 1979 are indicated. Percent ice cover calculations were made as outlined in the text using the area blocked out as indicated over the eastern Bering Sea (after Niebauer 1980).

DATA

Weekly mean southern ice limit data for the Bering Sea were obtained from the Naval Fleet Weather Facilities in Suitland, Maryland. The estimates represent both satellite and visual observations. The weekly percentage of ice coverage was calculated as the ratio of ice coverage to total area (ice plus open water) considered in Fig. 9-1. This gives a quanti-

tative estimate of ice coverage (hereafter called ice) but does not take into account ice thickness or concentrations (e.g., 1/8 concentration is weighted equally with 8/8 concentrations).

Monthly mean SSTs were obtained from the Naval Fleet Numerical Weather Central, Monterey, California, through D. R. McLain of National Marine Fisheries Service in Monterey. The data represent analyses of temperature reports made by ships of oppor-

Figure 9-2, a-f Percent ice cover as calculated from the area indicated in Fig. 9-1 for the eastern Bering Sea for the winters 1973-79. The shaded area is the six-year mean (seasonal cycle) while the solid line is the individual winter data (redrawn after Niebauer 1980).

tunity in the area. Bottom temperatures were obtained from Coachman and Charnell (1979). Northern Hemisphere 700 mb pressure charts were obtained from *Monthly Weather Review*. Air temperatures and surface winds for the Pribilof Islands were taken from the Local Climatological Data published by the U.S. Department of Commerce.

RESULTS

Seasonal ice cycle

Fig. 9-2 (a-f) illustrates the mean seasonal cycle of ice for the eastern Bering Sea. The ice generally begins its seasonal southward formation in November.

It is estimated that about 97 percent of the ice in the Bering Sea (Leonov 1960) is formed within the Bering Sea. Very little ice is transported south through the Bering Strait (Tabata 1974). The ice apparently forms like a giant conveyor belt, being generated along the south-facing coasts in the Bering Sea and moving southward at as much as 0.5 m/sec before finally melting at its southern limit (Pease, this volume). Therefore, although the ice on the Bering Sea shelf is subsequently pushed around by the wind on shorter time scales (see, for example, Muench and Ahls 1976), only about 3 percent of the seasonal sea ice cover is actually advected onto the shelf through the Bering Strait.

Seasonal ice formation progresses at an average rate of 12-13 percent per month of the area of the eastern shelf considered in Fig. 9-1, reaching 60-65 percent coverage in late March. The ice advance generally consists of a short, rapid advance (~ 24 percent per month) in November-December before slowing to ~ 6 -7 percent per month in December-March. With the exception of the rapid advance in November and part of December, the ice appears to dissipate faster than it forms, at about 18-20 percent per month in late March to early July. Lisitsyn (1960) has reported that during the period of ice retreat, 63 percent of the ice melts within the Bering Sea basin. The remainder leaves the Bering Sea by way of the various straits and passes (Fig. 9-1).

Interannual departures from seasonal ice cycles

Before we consider year-to-year departures from mean seasonal cycles, it is instructive to look at a longer time series of mean annual SST from the Pribilof Islands (Fig. 9-3). The annual SST was near the 15-year mean of 4.2 C in 1963 before rising to 5.4 C in 1967. SST then fell to 2.8 C, or 1.4 C below normal in 1975. Since then there has been a rapid rise, to 5.7 C, or 1.5 C above normal, in 1978. Similar data from Bristol Bay (Fig. 9-3) display similar characteristics. Thus, for the period of our interest (i.e., the last five years, 1974-79, of this time series) the mean annual SST of the southeastern Bering Sea shelf water has risen nearly 3 C.

Consider now departures from the seasonal means of ice (Fig. 9-2) and annual SST. A maximum in ice coverage and relatively low SST occur in the spring of 1976. For these reasons January 1976 has been arbitrarily chosen as a point at which to divide the time series into two sections. The first section is characterized by lower-than-normal sea temperatures and above-normal ice cover. This period coincides with above-normal upper (3,000 m or 700 mb) air flow from the north as outlined by McLain and Favorite (1976) and Niebauer (1980).

An extreme example of above-normal ice cover on the eastern shelf occurred in April of 1976 when the ice cover was ~ 25 percent above the mean. An example of the relatively strong upper air flow from the north during this period is shown in Fig. 9-4. Here a mean low is situated over eastern Siberia, causing southward flow off the Arctic Ocean down over Siberia and then eastward flow over the Bering Sea. Some indication of the severity of the conditions is given by the mean air temperature for April for the Pribilof Islands of -8.2 C (6.2 C below normal) under a mean northerly component of the surface winds of 2.2 m/sec (~ 0.4 m/sec above normal

Figure 9-3. Mean annual sea surface temperatures from the area of the Pribilof Islands and Bristol Bay and shelf bottom water temperatures (Coachman and Charnell 1979) for June for the S.E. Bering Sea (after Niebauer 1980).

northerly flow). In fact, for the period 1974-76 there were only five months in which there was monthly mean southerly flow and these were all summer months.

The second period (~ 1976 -79) is characterized by a strong rise in SST (~ 1 C/yr) and a precipitous fall (~ 9 -10 percent/yr) in ice (Niebauer 1980). These observations coincide and are probably a result of the abrupt swing of the upper air flow from northerly to southerly as reported by Niebauer (Chapter 3, this volume; 1980). Illustrated in Fig. 9-5 is an example of the strong flow from the south which probably caused the extreme below-normal (~ 35 percent below normal) ice extent in January 1979. This strong southerly flow persisted for most of the 1978-79 winter, resulting in over 40 percent below

Figure 9-4. 700-mb (~33 km) height pressure chart for April 1976 (labeled in dekameters) (after Wagner 1976).

normal ice cover by April 1979. Again, some indication of the mildness of conditions in the Pribilof Islands is given by the mean monthly temperature for April 1979 of 2.3 C (~4.3 C above normal) under the 2.3 m/sec southerly component of the surface wind (~3.9 m/sec above normal southerly flow). During this period of January 1976 to May 1979 there were 13 months during which there were mean southerly winds in the Pribilof Islands. Seven of these months were within the November-May ice season.

Fig. 9-6 illustrates the regression of the January southern ice extent over the years 1975-79, showing that the ice edge in the eastern Bering Sea has retreated ~550 km over this period. Table 9-1 shows that the rather dramatic rise in the mean January air temperature in the eastern Bering Sea coincided with the ice retreat and the shifting of the north-south component of the surface wind from north to south. Thus, in the period 1975-79 the mean January air temperature has risen ~9.3 C, coinciding with a northerly to southerly shift in winds of ~6.2 m/sec

(~12 kn) over 1976-79. As mentioned previously, this coincides with the dramatic retreat of the ice edge (Figs. 9-2 and 9-6) and the rise in SST (Fig. 9-3).

TABLE 9-1

January mean air temperatures with deviation from the mean and the monthly mean North-South component of the surface wind with the deviation from the mean from the Pribilof Islands for the period 1975-79

January	Air temperature (T°)	Deviation from mean (T° C)	N-S surface wind (m/sec)	Deviation from mean (m/sec)
1975	-8.0	-4.7	-1.8	-0.5
1976	5.6	2.3	3.8	2.5
1977	-0.8	2.5	-2.7	-1.4
1978	0.9	4.2	1.8	3.1
1979	1.3	4.6	2.4	3.7

Figure 9-5. Same as Fig. 9-4 except for January 1979 (after Wagner 1979).

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Figure 9-6. January southern ice limit for 1975-79.

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Remote Sensing Analysis of Ice Growth and Distribution in the Eastern Bering Sea

10

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ABSTRACT

Ice thickness distribution and ice types for the eastern Bering Sea are inferred from satellite imagery and available aircraft data. These have been combined with analyses of ice bridging and floe trajectories to estimate movement and generation of ice within the pack. The location of the ice edge has been plotted for different dates using satellite imagery and ice analysis charts. The position of the edge and the variability of the geographic location of the ice types and thicknesses supports a theory of ice generation according to which ice forms along the leeward side of east-west-trending coasts, is advected to the south-southwest within the pack, is broken into floes near the ice edge by the effects of wave propagation, and melts at the edge when the thermodynamic limits of its stability are reached.

INTRODUCTION

In March, 1979, a joint experiment was conducted by NOAA, NASA, and the University of Washington in the Bering Sea. The NOAA ship *Surveyor*, stationed along the ice edge, provided ground truth for physical oceanographic and meteorological experiments (Pease 1979). Personnel from the University of Washington tracked floe movement at the ice edge and obtained core samples of the ice (Martin and Bauer, this volume). Researchers from Scott Polar Research Institute installed accelerometers on floes to assess wave propagation into the pack. The NASA C130 aircraft flew one mission 14 March 1979 over the *Surveyor*. The remote sensing equipment on board included a laser profilometer, a step-frequency radiometer (Harrington et al. 1979), a scatterometer, and a 23-cm format, 153.12-mm focal length camera. The NASA C131 aircraft flew over the same location

a few hours later with a Side-looking Airborne Radar (SLAR) (Schertler 1979) on March 14. An additional track was flown in the Norton Sound area March 27. The *Surveyor* was in the ice during this period of maximum extent and at the beginning of a period of gradual retreat (Pease 1979).

This paper examines remote sensing evidence about the ice generation regime in the Bering Sea. Satellite imagery from the Defense Military Satellite Program, TIROS, and LANDSAT are used in conjunction with available ice charts from the Navy/NOAA Joint Ice Center (FWS) in Suitland, Maryland, and the above-mentioned field program, to formulate daily charts on ice conditions (Fig. 10-1). These charts were averaged to show weekly conditions. This data set is used to describe the ice regime in the eastern Bering Sea for March 1979.

Because ice in the Bering Sea is not a closed pack, as in the Arctic, the ice regime is qualitatively different. In the Bering Sea all ice is first-year ice and melts completely by the end of each season. The period 1-31 March 1979 was unique in that the ice was at maximum extent and also began a rapid, complete meltback to ice-free conditions.

WEEKLY AVERAGE ICE CONDITIONS

Four charts of weekly ice conditions have been produced by averaging daily charts. The charts cover the following periods: (1) March 1-7; (2) March 8-14; (3) March 15-21; (4) March 22-28. (Clouds obscured conditions at the very end of the month; extrapolations were not made.)

Figure 10-1. Daily charts on ice conditions in the Bering Sea. These analyses were compiled using Defense Military Satellite Program, TIROS, and LANDSAT imagery in comparison with FWS ice charts.

1. March 1-7 (Fig. 10-2). The ice is at maximum extent. Significant features include polynyas south of St. Lawrence Island and Nunivak Island and south of the western mainland in Norton Sound; floes which stream from the east of St. Lawrence Island to the south and southwest around the polynya area; the thick first-year ice (120 cm-2 m) north of St. Lawrence Island and Nunivak Island; and loose floes and streamers along the ice edge.

2. March 8-14 (Fig. 10-3). This week showed little change in ice conditions from the previous week; however, the ice continued to advance during the early part of the week and then began a gradual retreat.

3. March 15-21 (Fig. 10-4). A warming trend began during this week, accompanied by a reversal of northeast winds to a southerly direction (Pease, this volume). Breakup began and is especially noticeable in areas near the coast previously occupied by young first-year ice and polynyas.

4. March 22-28 (Fig. 10-5). This figure shows the effects of the breakup due to a warming trend. There is open water north of St. Lawrence Island and west of the mainland. The stream of floes around the south-southwest coast of St. Lawrence Island is still evident as are the floes and thicker ice west of the island.

The geographic constancy of features such as floe streamers, polynyas, and ice bands along the edge supports a theory of ice generation for the eastern Bering Sea like that which was proposed for the western Bering Sea by Loshchilov (1974) during the BESEX experiment. A schematic diagram (Fig. 10-6) shows that the ice, when at maximum extent, operates like a conveyor belt. The ice is formed in the leeward side of the east-west trending coasts and is advected downwind into the pack. This has also been observed by Muench and Ahlnäs (1976) and Ahlnäs and Wendler (1979). As the ice advances, it thickens and continues to be blown downwind to the edge, where it melts. This type of morphology for Bering Sea ice is evidenced by cores taken from the ice (Martin and Bauer, this volume, Martin and Kauffman 1979, Gloerson et al. 1974), which show ice compaction north of the islands and divergence around the islands and along the edge of the pack. The ice in the area to the east of St. Lawrence Island changes gradually from compaction to divergence. This is due, in part, to the shape of the ice pack, which is formed in a more constricted area than the one to which it is advected. Remote sensing data collected in March support the hypothesis that (1) ice is formed in the leeward side of east-west trending coasts under a dominant wind regime from the northeast; (2) ice is compacted north of the islands; (3) ice diverges around the islands; (4) ice at the seaward edge has reached its thermodynamic limit and is in the process of decay; and (5) ice retreat during breakup progresses most rapidly into the areas of thinner ice which had been supplying ice to the rest of the pack.

Figure 10-2. Weekly average ice conditions, 1-7 March 1979.

Figure 10-3. Weekly average ice conditions, 8-15 March 1979.

Figure 10-4. Weekly average ice conditions, 15-21 March 1979.

Figure 10-5. Weekly average ice conditions, 22-28 March 1979.

Figure 10-6. Schematic diagram of areas of ice generation, compaction, and divergence. This is meant as a first look for future refinement. The area of Eastern Norton Sound under certain wind conditions also behaves as an ice-formation area.

ICE FORMATION

In the initial stages of ice formation, grease ice is produced in the upper layer of the water and floats to the surface (Martin, *Oil-Ice Interaction*, this volume). This ice is blown downwind on the water's surface and piles up at the leading edge of ice streamers (Fig. 10-7). As the grease ice layer thickens, it damps the surface wave field (Martin, *Oil-Ice Interaction*, this volume) and is compacted until it forms a surface layer which can support ice growth underneath. This thickening ice forms pancakes and eventually larger floes which are incorporated into the pack. This replacement process appears to be continuous.

Fig. 10-8 shows a LANDSAT 3 image of the polynya area west of Nome along the Seward Peninsula. It has been enhanced to bring out detail in the first two gray levels of the image so that the grease ice forming in the lee of the mainland is visible. The

wind was from the northeast (Pease, this volume). Streamers of grease ice can be seen, as well as large floes made of thin ice which was piled up and had broken free. Fig. 10-9, a mosaic of five LANDSAT 3 scenes from 12 March 1979, shows ice formation along a track from the Bering Strait to the ice edge near St. Matthew Island. Here grease ice is visible south of St. Lawrence Island. The gray signature of this thinner ice gradually appears whiter due to the thicker ice downwind. This polynya with grease ice forming within it was also noted during BESEX in 1973 (Ramseier et al. 1974), and is often evident in FWS ice charts (Eastern-Western Arctic Sea Ice Analyses, 1972-78). The wind was from the north (Pease, this volume).

Cores taken by Martin and Kauffman (1979) at the ice edge show that the upper portion of these floes consisted of a layer of consolidated grease ice. This was also observed in cores taken during BESEX in 1973 (Gloerson et al. 1974, Ramseier et al. 1974).

Figure 10-7. Grease ice photo taken from the CV990 aircraft during BESEX, 1973. (Courtesy of S. Martin, Univ. of Washington, NASA/Goddard). The ice is being blown downwind where it piles up at the leading edge. The grease ice streaks also damp the waves at the surface.

Figure 10-8. 2 March 1979, LANDSAT Imagery. This image has been enhanced to show details of the grease-ice formations in the leeward side of the shoreline.

These photographs support the hypothesis that the ice is formed to the north in large polynya areas and is blown south by the wind to the ice edge.

ICE COMPACTION

Ice in the Bering Sea, moving under the influence of the prevailing wind, tends to compact as it constricts or meets an obstacle. Sodhi (1977) and Shapiro and Burns (1975) show that this type of ice bridging often occurs north of St. Lawrence Island. FWS ice charts often show areas of thick first-year ice north of St. Lawrence, Nunivak, and St. Matthew islands (Eastern-Western Arctic Sea Ice Analyses 1972-78). Fig. 10-9 shows an example of such ice compaction. The wind is from the north and ice on

the north side of St. Lawrence and St. Matthew Islands appears to be thicker, consistently white in appearance, and composed of tightly compacted floes with ridging structure evident on the surface. On the next day, 13 March, the LANDSAT 3 image (Fig. 10-10) shows that the same area of compaction north of St. Lawrence Island is relatively unchanged. The ice on both days was at maximum extent (Fig. 10-3).

The same phenomenon can be seen earlier in the month in the 2 March imagery from LANDSAT 3 (Fig. 10-11). The gray scale of the image differs significantly from the previous image because less light was available. However, the same areas of relative thickness can be seen north of St. Lawrence Island. The wind was from the northeast (Pease, this volume).

Figure 10-9. 12 March 1979, LANDSAT. This image is a mosaic of five separate scenes. It shows convergence north of St. Lawrence Island, St. Matthew Island, the polynya behind Lawrence Island, breakout of floes through Bering Strait, floes moving around St. Lawrence Island, ice streamers and bands at the edge, and roll cloud formation.

Figure 10-10. 13 March 1979, LANDSAT.

ICE DIVERGENCE

Ice which is compacting tends to raft and ridge until shearing causes portions of the ice to break off. The freed ice then separates from these bridging areas, causing leads to be formed (Sodhi 1977). This ice moves downwind as giant floes and vast floes in a matrix of smaller floes and brash ice (an accumulation of floating ice made of fragments <2 m across).

This accounts for the areas with large floe streams in Figs. 10-1-10-5. This phenomenon shows up well in both Fig. 10-9 and 10-10, and was noted during the BESEX experiment by Ramseier et al. (1974), Gloerson et al. (1974), and Campbell et al. (1974). When locations of the leads and floes for both days are compared, it is seen that the floes have moved, and that the orientation of the leads remained consistent, approximately perpendicular to the wind.

Figure 10-11. 16 March 1979, TIROS. This is an enlargement of a portion of the TIROS scene. Enlargements like this were used to track floes for 16, 17, 18, and 19 March.

A more detailed study of this phenomenon was undertaken using enhanced TIROS imagery for March 16, 17, 18, and 19 (Fig. 10-12), so that individual floes could be tracked for a longer period of time. Muench and Ahlñäs (1976) tracked floes during spring 1974 and found that their dominant direction of travel, in the area south of St. Lawrence Island, was to the south-southwest under predominantly northeast winds. Twelve floes A-G, J-N (Fig. 10-12), were tracked during the period 16-19 March for two days, and where possible for three days. Due to the resolution of the imagery, all floes appeared to be giant (>10 km across). Floes could not be tracked

longer than three days because they tended to break up into floes smaller than the resolution of the imagery. Table 10-1 lists these floes by letter, gives their approximate size, the direction of the wind from true north (Pease, this volume), the floe direction, and the floe speed.

Floes A-G were followed on 16-17 March. For B-F the wind direction was estimated at 220° - 225° relative to true north. The floes traveled 235° to 245° . For A and G the winds were estimated to be at 205° and the floes traveled 180° . These floes are within an area where their movement is affected by the ice shear around St. Lawrence Island. For floes

Figure 10-12. Floe trajectory map, 16-19 March 1979. The dominant wind regime was from the northeast during this period; floes tended to move to the south-southwest in response to the wind around the polynya area behind St. Lawrence Island.

TABLE 10-1
Floe trajectories, March 16 - 19, 1979

Floe	Wind direction (from true north)	Floe direction	Floe velocity (m/sec)	Floe size (km)
A	205°	180°	.28	8×20
B	220°-230°	256°	.13	16×40
	220°-230°	218°	.50	16×40
C	220°-225°	252°	.23	12×20
	220°-225°	238°	.23	12×20
D	220°-225°	254°	.23	16×20
E	220°-225°	237°	.22	20×20
F	220°-225°	245°	.22	36×32
G	205°	180°	.22	12×12
J	230°-235°	238°	.28	16×36
K	230°-235°	238°	.46	30×28
L	230°-235°	230°	.41	16×32
M	230°-235°	235°	.39	14×34
N	210°	217°	.42	32×60

J-M, on the 18th and 19th, the winds were from 230° to 235° and the floes traveled 230°-238°. For N, the floe traveled 217° under winds estimated to be at 210°. All floe movements show a strong correlation with wind direction and appear to move to the south-southwest under the northeast wind regime which is dominant at that time of year (OCSEAP 1977). The ice movement is slightly to the right of the wind. This agrees with data reported by Sverdrup (1928). Floes B and C were tracked for three days and maintained the strong relationship between wind direction and direction of floe movement.

The average distance traveled by the floes during a 24-hour period was 26 km with a range of speed from 0.13 m/sec to 0.50 m/sec. The average speed was 0.26 m/sec. The velocity, distance, and direction of travel of these floes imply a strong divergence of ice within the pack downwind toward the ice edge. The lead orientation for the same period (Fig. 10-13) reflects the movement of the ice to the southwest under this wind pattern and suggests a movement of the ice toward the edge to the west of St. Matthew Island. SLAR data from the NASA C131 aircraft (Fig. 10-14) also show this lead pattern east of St. Lawrence Island in an area of large floes. Winds this day were also from the northeast (Pease, this volume).

ICE-EDGE PHENOMENA

Figs. 10-2 and 10-3 show little change in the location of the edge itself, yet floe studies indicate that volumes of ice from within the pack are being blown to the edge. In order to study the nature of the ice at the edge, ground truth data needed to be collected.

The NOAA ship *Surveyor* was stationed at the ice edge from 2 to 14 March 1979 (Pease 1979; Pease, this volume). On 14 March, two NASA aircraft flew over the ship while ground truth data were being collected. The flightline for the C131 is shown in Fig. 10-14. Fig. 10-15 shows the flight track from the C130 aircraft. An analysis of *in-situ* data on ice cores (Martin and Kauffman 1979) and floe drift, along with information on wave attenuation, shows that ice along the edge consists of rotten floes which have reached a thermodynamic limit of stability and have begun to melt (Martin and Bauer, Pease, this volume). A plot of the location of the isotherms of the water surface as observed during the cruise period shows that the -1 C isotherm moved southward with respect to the ice edge (Pease, this volume). The CTD casts taken also show a lens of less saline water extending out from the edge—evidence of melting (Pease, this volume).

Figure 10-13. Lead orientation, 16-19 March 1979. Leads follow the same general trend downwind as the floes.

Satellite imagery such as that seen in Fig. 10-11, TIROS, and Figs. 10-9 and 10-10, LANDSAT, were used to study small-scale variations of these ice bands. These band features are not an isolated occurrence. They are often noted on FWS charts and have been studied by Muench and Charnell (1977). They can also be seen in a LANDSAT 3 image taken 5 March 1979 (Fig. 10-16), and on the 14 March imagery (Fig. 10-14).

An assessment of change in an ice band can be made using a photograph from the C130 overflight (Fig. 10-17). Although this image was taken approximately three hours earlier than the C131 SLAR image, the basic shape of the formation is still recognizable and can be seen to be made up of small, angular floes.

Similar photography from successive altitudes was used to study the characteristics of the floes within these bands. Fig. 10-18 shows a cross section of one of these bands. The distance across the band is 20 km. Darker floes in the middle of the band indicate thinner ice. The darker color and the nature of the surface patterns suggest that these floes are rotting. Ship's personnel who had occupied stations on similar floes reported that this type of ice was melting rapidly, and that it was thin enough to respond plastically to waves which were propagating through the ice (Pease 1979; Martin and Bauer, this volume).

Fig. 10-19, taken from a lower altitude, shows another cross section of a band. The flightline was from west to east, the wind from the northeast.

SLAR IMAGERY

MARCH 14, 1979

BERING SEA

Imagery from NASA
Lewis Research Center

Figure 10-14. 14 March 1979, SLAR Mosaic. (Courtesy of R. Schertler, NASA/Lewis.) The C131 flightline is on the map at the middle right. Only the ice-edge lines and a portion of the long line from Nome are reproduced.

Figure 10-15. 14 March 1979, C130 flightline.

Smaller floes were blown downwind to the ice band. The leading edge of the band is made up of angular, thicker white ice floes. These floes are held together at the leading edge by the effects of the incoming swell; yet, as a group, they tend to move downwind at speeds of 0.3 m/sec (Martin and Bauer, this volume).

Fig. 10-20 shows the surface configuration of melting found on one of the thinner floes. This melting is not like that of Arctic ice, where melt puddles form on the surface of thick floes, but appears to be due to floe motions causing water to spill over the sides, and to the effects of a surrounding area of warmer water which causes the floe to rot. The two basic types of floes found at the edge were the highly rafted, relatively thick, white, angular floes (Fig. 10-21), whose freeboard allowed them to travel rapidly under the influence of the wind; and the thinner, larger floes (Fig. 10-22) found in the interior of the bands. These larger floes were variable in size,

but were generally an order of magnitude larger than the white floes, which typically measured between 30 and 60 m in diameter. These sizes compare well with those measured by Martin and Kauffman (1979).

Martin and Bauer (this volume) have hypothesized that floes along the ice edge are formed from pack ice which is broken up at the edge by the effects of wind and incoming swell. First, the sheet ice is broken into vast rectangular floes whose width is proportional to the wavelength of the incoming swell (for an illustration, see Fig. 12-14, Martin and Bauer, this volume). These floes are further broken down and the thicker ice is rafted while being broken into small floes, which have sharp, angular edges, unlike pancake ice, which tends to be rounded. Pease (1979, this volume) noted from CTD casts along the ice edge that there is a surface lens of less saline, warmer water extending from the ice edge—evidence of meltwater. Preliminary returns analysed from the step-frequency radiometer on board the C130 tend to support

Figure 10-16. 5 March 1979, LANDSAT. This is a two-image mosaic showing ice bands to the southeast of Nunivak Island.

Figure 10-17. 14 March 1979, C130 photo mosaic of an ice band also seen in the 14 March SLAR data run.

Figure 10-18. Ice band near the NOAA ship *Surveyor*, showing the two types of ice in a band approximately 20 km across. The light-colored, angular floes are thicker ice; the darker ice near the center is thinner, melting ice.

Figure 10-19. Ice band near the NOAA ship *Surveyor*. The flightline was from west to east (left to right), the wind was from the northeast. Ice fragments can be seen to be advecting downwind to the band. The leading edge of the band is loosely held together by the influence of incoming swell.

Figure 10-20. Two-frame mosaic of thinner ice near the interior of an ice band. The floes show the effects of melting along the edges and in the center, especially.

Figure 10-21. Detail of thicker, rafted white floes. Note the angular edges and the presence of snow on the surface. The floes are 30-60 m in size.

this hypothesis. The return from the radiometer indicates that the ice is relatively fresh and occurs as a matrix of less-saline water than would normally be found on the surface (Swift, personal communication).

ICE RETREAT

At the end of March 1979, the ice in the Bering Sea began a rapid retreat. This can be seen rather dramatically by comparing the 16 March TIROS

Figure 10-22. Detail of thin, melting floes. Note the lack of snow on the surface and thaw holes.

image (Fig. 10-11) with the 26 March TIROS image (Fig. 10-23). This period of retreat coincided with a time of maximum insolation and warm southerly winds generated by a low over the Aleutians (Pease, this volume). A comparison of Fig. 10-23 with the average weekly chart for 22-28 March (Fig. 10-5) shows the meltback proceeding in the eastern Bering Sea between the coast and Nunivak and St. Lawrence

Islands, an area into which ice is continually advected (Fig. 10-6). It is interesting to note the persistence of the floes around St. Lawrence Island. The SLAR data for March 27 (Fig. 10-24) shows an open water/thin ice area north of St. Lawrence Island (McNutt 1977), more leads present on the eastern edge of St. Lawrence Island, and open areas and new leads in Norton Sound.

Figure 10-23. 26 March 1979. This TIROS enlargement shows the effects of a rapid meltback. (Compare with Fig. 10-11.)

SLAR IMAGERY MARCH 27, 1979 NORTON SOUND

Imagery from NASA Lewis Research Center

Figure 10-24. 27 March 1979. SLAR mosaic of a flightline in Norton Sound near St. Lawrence Island. The percentage of open water is increasing, and an open-water area is evident north of St. Lawrence Island. (Courtesy of R. Schertler, NASA/Lewis.)

The ice during this period of meltback appeared to be retreating most rapidly in areas which had previously been supplied by ice blown down from the north. The change of wind direction from north-northeast to south-southeast affects floe trajectories and tends to blow ice to the north instead of to the south-southwest. One would not expect to find ice bands along the edge under these conditions, and yet isolated, string-like features of ice persist in areas which are otherwise ice free (Fig. 10-23). In some areas, e.g., the area to the southwest of St. Lawrence Island, ice persists and remains relatively motionless. More work needs to be done to study the effects of wind, insolation, and water temperature on ice movement during this retreat.

CONCLUSIONS

The 1978-79 ice season was a relatively light ice year, perhaps because of short-term climatic variations in the mean annual sea surface temperatures (Niebauer, this volume). March 1979 was unique, in that one month encompassed conditions of maximum extent and rapid retreat. When the Bering Sea ice is at maximum extent, it reaches a steady state in which ice is formed in polynyas on the leeward side of east-west trending coasts and advected downwind. The ice bridges along the northern coasts of islands and shear zones existing on both sides of the wedge of thicker ice. Floes are broken off at these shear zones and, under north-northeast wind conditions, floes originating on the eastern side of St. Lawrence Island drift south-southwest towards the ice edge near St. Matthew Island. Ice at the edge is broken up by the effects of wind and incoming swell and is blown downwind into ice bands. Ice in these bands reaches its thermodynamic limit and melts along the edge. During the breakup the ice retreats under the effects of increased insolation and southerly winds. The ice near the coast, which had been diverging from the north, is no longer replenished and retreats first. Under continued southerly wind conditions, individual floes may begin to migrate northward.

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Nearshore Ice Characteristics in the Eastern Bering Sea

11

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ABSTRACT

Bering Sea nearshore ice conditions are described on the basis of a compilation of fast-ice edge satellite data, observations of specific ice events, and results from other studies. LANDSAT imagery at 1:500,000 scale was used to map Bering Sea ice conditions between 1973 and 1976 in nearshore areas. From these maps, secondary single-attribute maps were compiled, giving the edge of fast ice at various epochs during these four years. Seasonal maps representing midwinter, late winter to early spring, and mid-to-late spring were then compiled. The seasonal average maps were then compared to determine seasonal trends in the location of the fast-ice edge.

This information was analyzed together with imagery showing specific ice events and bathymetric charts, wind data, tidal variations, and observed ice trajectories. The result is a regional description of average nearshore ice conditions along the Bering Sea coast from Cape Prince of Wales to Cold Bay on the Alaska Peninsula. Over this distance a north-south transition is found from fast-ice conditions similar to those in the Beaufort and Chukchi Seas (fast ice bounded by grounded ridge systems at a depth of 20 m) to conditions generated by large tidal variations, offshore winds, and highly mobile ice, with the result that fast ice is found only in highly protected, shallow waters.

INTRODUCTION

Any discussion of oceanic ice in the nearshore area necessarily centers on fast ice—ice fixed with respect to shore. Although fast ice is found along almost all ice-bound coasts, its characteristics vary from one place to another. This variation depends on many factors, including the local bathymetry, internal stresses in the adjacent ice pack, local surface winds, tides, and currents. Usually the fast ice is composed largely of annual ice, with perhaps occasional interfused pieces of multiyear ice. In Alaskan waters, annual ice seldom grows to a thickness of more than two meters. Because of the low buoyancy of ice, most of the vertical extent of fast ice is below the surface level. Obviously then, at water depths of less than two meters, the fast ice is actually bottom-fast after it grows thick enough.

Changes in sea level, resulting either from tides or from weather patterns, create a “hinge” between the floating fast ice and the bottom-fast ice. The hinge usually takes the form of a crack between the two ice types. As the winter progresses and the ice grows in thickness, the active tidal crack will generally move seaward, leaving old cracks to the shoreward, often bridged by blowing snow. In areas where tidal variations are low, the pattern of these tidal cracks can be fairly simple. Where there are large tidal variations, as in the Bering Sea, there will be a tidal crack *zone* with the currently active tidal cracks determined largely by the instantaneous tide state as well as ice thickness.

This pattern is superimposed on the ice state created during freeze-up in the nearshore area. It is possible for the ice to simply freeze in place, growing thicker with time, but this is often not the case. In reality a wide variety of ice conditions can be found, depending on the history of ice dynamics in a particular freeze-up season. The original ice sheet, for instance, may freeze to a thickness of 30 cm; a storm may ensue that withdraws the ice from shore, breaks most of it up into small (1 m) plates, and then drives it to shore again. The plates may then freeze together in an extensive rubble field and form the fast ice for that year.

Other initial conditions are possible. In 1973, an “ice push” event occurred at Kotzebue, Alaska: a stable sheet of moderately thick (1-2 m) fast ice was driven as much as 15 m onto the beach just south of town, carrying with it a surplus landing barge used as a salmon cannery (Mr. Albert Francis, personal communication). Kovacs and Sodhi (1979) have documented a number of these incidents in the Beaufort Sea region, as well as a related phenomenon, ice piling events: instead of an ice sheet being pushed across the beach and adjoining tundra, a large pile of broken ice is created at or near the beach.

Figure 11-1. Idealized Beaufort Sea nearshore ice regime showing one of many possible configurations. Note apron of new ice adjacent to floating shear ridge in attached fast ice zone. This ice has formed as a result of a recent advection event.

Offshore, the floating fast ice is often anchored by pressure and shear ridges with sufficient keel depth to be grounded on the ocean floor. Generally, few large grounded ridges are found in shallow water (up to 12 m) and ridges are seldom thick enough to be grounded in water deeper than 20 m. While a great deal of work has been done toward determining these limits in the Beaufort Sea (Reimnitz and Barnes 1973, Kovacs and Mellor 1974, Kovacs 1976, and Stringer 1978), relatively little work has been done in the Bering (Stringer 1978, Ray and Dupré, this volume, Dupré 1980).

Floating fast ice is not always bounded by a zone of grounded ridges (sometimes called *stamukhi*, see Reimnitz and Barnes 1976). Whether or not a grounded-ridge zone exits, fast ice can extend seaward up to 100 km or more (Stringer 1974). If the grounded-ridge zone is present, it tends to protect the enclosed fast ice from deformation resulting from pack-ice forces, although deformations may still take place within this protected zone.

The grounded-ridge (or *stamukhi*) zone is an important feature because it often determines the boundary between fast ice and pack ice. The "flaw

Figure 11-2. Idealized Beaufort Sea nearshore ice regime showing one of many possible configurations. Note pile of ice on beach, large apron of attached ice starting at a shear ridge and extending seaward.

lead" is often found just seaward of the deepest grounded ridge. In this zone a large amount of pack-ice energy is expended that must be accounted for when modeling nearshore ice mechanics. With the increased attention to offshore structures required in the course of petroleum development, the grounded-ridge zone has become important in relation to physical hazards.

Beyond the grounded-ridge zone, an apron of floating fast ice (here called "attached" ice in order to emphasize the absence of grounded features to the seaward) can often be found extending from a

few meters to many kilometers into the ocean. Since the stability of attached ice is tenuous, it can easily be converted into pack ice by an ice-breaking event.

Figs 11-1 and 11-2 give some idea of the range of conditions that can be found. The situation depicted in Fig. 11-1 shows relatively undeformed bottom-fast ice along the beach with tidal cracks occurring near the 2-m isobath. Offshore in water a few meters deep, occasional piles of pressured ice may in fact be grounded. These often act as single-point anchors and are generally created as weaker ice is pressured around stronger pans. This pattern extends out to

the grounded ridges. The dimensions of the pans and piles around them, as well as the distance to the grounded ridges, can vary widely. For instance, the pans could be 30 or 3,000 m in diameter and the ice piles could be 1 or 10 m above sea level. The distance to the grounded ridges could be from 1 to 30 km. Because of the necessary vertical exaggeration in these figures, the angle of repose of ice ridges shown appears to be much steeper than it is. Furthermore, the thickness of unpressured ice is exaggerated in the vertical plane, which may give a false impression of the geometry involved.

Beyond the grounded ridges, the attached ice is depicted as being relatively smooth but with some hummocking. Finally a large floating ridge is encountered which would be grounded if it were further inshore. As depicted here, this ridge was recently the edge of fast ice with active differential motion taking place along it. But the ice opened and moved seaward, forming a large flaw lead, which froze to a thickness of 10 or 20 cm and then failed in tension, forming a new flaw lead. This narrow lead now defined the edge of fast ice. Beyond the lead, the ice can truly be classified as pack ice.

Fig. 11-2 shows another common nearshore ice situation. Here, a rubble pile is found on the beach with the active tidal crack beyond its base. Just offshore, ice is piled against the beach and grounded in a few places. In some years many such ice piles are found near the beach in exposed locations (Barrow, for instance). Ice of this type makes activities involving transportation across the fast ice particularly difficult. Beyond the grounded ice, a second tidal crack may be found, followed by floating fast ice with occasional minor ridges which may be grounded in relatively shallow water. Farther offshore, there are smooth pans separated by hummocked ice and, finally, near the 14-m isobath is the grounded-ridge zone. As depicted here, the pack ice in the past has been driven along the outside edge of the outermost grounded ridge, creating a shear ridge similar to the floating ridge in the attached ice zone of Fig. 11-1. Here, however, the attached ice has not recently withdrawn but has remained adjacent to the grounded ridges, creating a zone of thin ice. The flaw lead is found offshore from a large floating ridge in the attached ice.

The pack ice begins at this point and extends seaward. Again, it should be emphasized that the vertical scale creates an inaccurate impression of horizontal dimensions: the distance between the grounded-ridge zone and the large offshore ridge could be on the order of 10 or 20 km. However, observations by LANDSAT (Stringer 1978) and laser

profilometer (Tucker et al. 1980) show that these large shear ridges formed by differential motions of ice sheets are generally found in the nearshore area. Hence, their frequency diminishes with increasing distance from the grounded-ridge zone.

FACTORS DETERMINING BERING SEA NEARSHORE ICE CONDITIONS

The description of nearshore ice conditions and behavior presented up to this point is largely based on observations in the Beaufort and Chukchi Seas and applies to some areas of the Bering Sea. However, two factors influence ice behavior in some areas of the Bering Sea that are almost totally absent in the Beaufort: tides and ice advection. While the Beaufort coast experiences tides with a variation of only a few decimeters, tides at many locations on the Bering coast range over several meters. Also, while Beaufort Sea ice is almost always packed in against the coast, in many places along the Bering coast the ice is almost continually being pushed away from shore by winds and currents (Stringer 1978, McNutt, Pease, Ray and Dupré, this volume).

Fig. 11-3 shows an ice profile more typical of fast ice in the Bering Sea than Figs. 11-1 and 11-2. A grounded ridge is shown some distance from shore, but certainly closer to shore than the 20-m isobath. In order to be even semipermanent, this ridge must be sufficiently grounded to withstand the buoyant forces during high tides, and must have such a geometry that tidal fluctuations will not cause disintegration. Obviously, grounded ridges cannot present a continuous dam against the large forces created during tidal variations; hence, breaks and other disruptions of these ridges are common.

Inshore from the grounded ridges, floating and bottom-fast ice are found. The extent of both these ice types depends greatly on the tide state, since the Bering coast has extensive mud flats covered by very shallow water. Several active and inactive tidal cracks can be found. Because of lateral motions caused by tidal currents and disruptions of the grounded ridges by tidal fluctuations, fast ice in the Bering Sea is not nearly so stable as fast ice in other areas with little tidal variation.

Attached ice can occasionally be found beyond the grounded-ridge system, but because of tidal variations, the flaw lead is most often found just seaward of the grounded ridges. Again, as in Figs. 11-1 and 11-2, the necessary vertical exaggeration should be considered when viewing this schematic drawing.

Figure 11-3. Idealized Bering Sea nearshore ice regime showing effect of large tidal variations on fast ice. As depicted here, low tide has caused the fracturing of floating fast ice and the breakup of attached fast ice. This effect tends to limit the extent and stability of Bering Sea fast ice.

Advective export of Bering Sea nearshore ice also contributes to the limitation of fast ice. There are many areas along the Bering coast where ice motion has a significant seaward component, and as a result, grounded ridges are seldom built in these locations. This condition contrasts sharply with that of nearshore ice in the Beaufort Sea, where the pack ice is nearly always present along the fast-ice boundary and is often driven along the fast ice with a shoreward

component of force, thus creating the well-known shear ridges often found in that area.

These two distinguishing influences, tidal fluctuations and ice advection, affect the Bering coast in varying degrees. At some locations they combine to severely limit the edge of fast ice to isobaths even less than 6 m. At other locations, conditions similar to those found in the Beaufort Sea prevail, with ice ridges grounded along the 20-m isobath.

THE LOCATION OF BERING SEA FAST ICE

In order to identify ice characteristics on a site-specific basis, maps of nearshore ice conditions were prepared from LANDSAT imagery at a scale of 1:500,000, showing the location of fast ice, pack ice, leads, ridges, hummock fields, and other identifiable features as well as shoreline and bathymetry. The techniques involved in preparing these maps have been described elsewhere (Stringer 1979) and will not be elaborated here.

The individual maps of LANDSAT scenes, each covering an area of approximately 160 km × 160 km (100 × 100 nautical miles), were reduced to a scale of 1:1,000,000 and combined to produce composite single-attribute maps of the Bering Sea nearshore area at specific instances. The most important characteristic of sea ice for determining nearshore ice conditions was found to be the edge of fast ice. The ice season was divided into three periods, and a series of three maps was compiled showing the location of the edge of fast ice at specific times over several years.

Figs. 11-4, 11-5, and 11-6 show fast-ice edges monitored during the periods of mid-winter, late winter to early spring, and mid-to-late spring between 1973 and 1976. Shown along with these ice edges is the average ice edge. In order to reveal temporal changes of location of the average ice edge, these average edges were compiled on one map (Fig. 11-7). Obviously, any significance accorded to trends apparent on this map must be tempered by consideration of the variability exhibited in the ice-edge data. At some locations, the edge of fast ice varies considerably in position during each period. Although the average edges in these locations show a temporal trend, it has only minor significance. In other locations, the variability of the fast-ice edge for each period is small compared to the changes in the average position from period to period. Strong temporal trends with year-to-year dependability are indicated.

Finally, Fig. 11-8 shows the regional nearshore ice characteristics of the Bering Sea. This map was prepared on the basis of the considerations just described and by analysis of LANDSAT images for ice features such as ridges and hummock fields and of smaller scale satellite imagery for dynamic ice motions and other nearshore characteristics. The following section adds more detail to the nearshore ice descriptions of Fig. 11-8 (items 1-33).

CHARACTERISTICS OF NEARSHORE ICE IN THE BERING SEA

In order to describe ice conditions along a segment of the coast, it is necessary first to describe the pack-ice conditions just offshore; the order of presentation here conforms to that necessity.

Ice motion through the Bering Strait

The Bering Strait, with the Diomed Islands in its center, represents a highly restricted ice passage between the Chukchi and Bering Seas. For some time, popular belief held that ice motion through the strait was generally from south to north in response to oceanic currents. This was supported by many ship-based springtime observations of broken pack ice passing northbound through the strait, often at speeds of several knots because late season lows to the south were causing southerly winds (Pease, this volume). Shapiro and Burns (1975) and Ahlnäs and Wendler (1979) have shown that occasional "break-out" events occur to the south when Chukchi Sea pack ice is extruded through the strait at fairly high velocities. These events have been examined in more detail by Prichard et al. (1979) and found to occur only during somewhat rare (two to three times annually) meteorological conditions which result in northerly winds across the ice in the strait, combined with a reversal of the usual south-to-north currents. During these events fairly large quantities of ice can pass through the strait, producing extensive grounded ice-ridge systems on Prince of Wales Shoal just to the north of the Alaskan side of the strait (Stringer 1978). It seems reasonable that during this process, relatively deep-draft first-year ice features would be created and transported into the Bering Sea and perhaps into western Norton Sound, but not eastern Norton Sound (Ahlnäs and Wendler 1979).

Nearshore ice conditions—Bering Strait to Yukon Delta

At Wales, the edge of fast ice closely follows the coast, largely as a result of the combined effect of deep waters (>20 m) adjacent to the coast and ice motions through Bering Strait. From Wales to the Port Clarence entrance, the edge of fast ice is generally inshore from the 20-m isobath in all periods.

At the entrance to Port Clarence, the edge of fast ice usually bridges the narrow embayment of the 20-m isobath into the port. However, a tongue of open water occasionally follows this indentation. Just south of the Port Clarence entrance lies a group of shoals as shallow as 4 m. These often appear to be

Figure 11-4. Map showing the location of the Bering Sea fast-ice edge during midwinter for 1974, 1975, and 1976.

Figure 11-5. Map showing the location of Bering Sea fast-ice edge during late winter/early spring for 1973 and 1974.

Figure 11-6. Map showing the location of Bering Sea fast-ice edge during mid-to-late spring for 1973, 1974, 1975, and 1976.

Figure 11-7. Map comparing average Bering Sea fast-ice edges for winter, late winter/early spring, and mid-to-late spring in order to determine seasonal changes.

Figure 11-8. Map summarizing Bering Sea regional nearshore ice characteristics.

anchoring the fast ice, sometimes creating a seaward bulge extending as far as the 20-m isobath. South of the Port Clarence shoals, the period edges of fast ice draw together and approach the 20-m isobath from the coastal side.

At Sledge Island the 20-m isobath makes an abrupt 80° turn to the east, entering Norton Sound. The period-average edges of fast ice follow this isobath past Nome and on to Cape Nome. In this region the edge of fast ice exhibits considerable variability during at least two periods. Hence, the agreement of the period averages should not lead to the conclusion that the edge of fast ice here is stable: in fact, it can occasionally be found well toward shore or beyond the 20-m isobath. Extensive grounded ice piles were observed off Nome during the spring of 1973, well inshore of the 20-m isobath. At that time, attached fast ice extended considerably farther offshore to approximately the 20-m isobath. Although ice is generally moving away from this coast, these ice piles are evidence that at times the ice can be driven against this shore.

Inside Norton Sound there is no correlation between the edge of fast ice and the 20-m isobath. From Cape Nome to Cape Darby, the edge of fast ice follows the coast rather closely in waters around 16 m deep. At Golovin Bay, just west of Cape Darby, the edge of fast ice shows considerable variability as it bridges the mouth of the bay.

At Cape Darby, the edge of fast ice follows the headland closely, making a 90° turn from a north-west-southeast trend to a northeast-southwest trend. The bathymetric configuration here is steep and the possibility of grounded features anchoring the ice decreases rapidly with distance from shore. From Cape Darby, the edge of fast ice is indented at the mouth of Norton Bay, nearly touching the headland at Cape Denbigh. This edge of fast ice follows more or less the 12-m isobath, but as the mid-winter ice-edge map (Fig. 11-4) shows, the edge of fast ice can exhibit some variability across Norton Bay.

From Cape Denbigh to Stuart Island the edge of fast ice characteristically bridges the southeastern end of Norton Sound, as far as 50 km from shore in waters very close to 20 m deep. However, the bottom of Norton Sound is relatively flat and it is unlikely that depth is a controlling factor in the location of the fast-ice edge here. Fig. 11-7 indicates a systematic trend in location of the average ice edge toward shore with time in this portion of Norton Sound. Analysis of the distribution in each period shows that this trend is reasonably valid. Furthermore, the variation from period to period for each year also shows this effect. However, the variation

within each period is such that only the midwinter average location of the fast-ice edge can be relied upon with any confidence. The late spring ice edge, for instance, can vary from near the shore to nearly 30 km seaward.

From Stuart Island, the fast-ice edge follows a westerly course to a point south of Cape Nome, yet 50-60 km offshore from the mouth of the Yukon River. At that point, it makes a broad turn to follow the coast to the south. The seaward extent of fast ice here appears to be determined largely by the location of the Yukon prodelta. The fast-ice edges are all in the vicinity of an underwater slope between two relatively flat plains at depths of 6 and 12 m. Fig. 11-7 shows that the average location of the fast-ice edge here builds seaward from midwinter to late winter to early spring, then erodes back even further than midwinter during the mid-to-late spring. This pattern appears to be caused by ice piling on the prodelta slope. Although many major ridges can be seen on LANDSAT imagery of the Beaufort Sea, only a few can be seen on Bering Sea images. This portion of the prodelta region is one of the places where such major ridges have been observed. Furthermore, Thor and Nelson (this volume) report a high density of ice-scour features in this area as a result of ridge-keel motions.

There is shoal at the very northwestern tip of the prodelta at a reported depth of 7 m. Schertler (1978) has reported floebergs imaged on side-looking airborne radar and in aerial photographs. Examination of the data suggests that these are rubble piles grounded on this shoal. LANDSAT imagery of this area from 1973 to the present indicates that it is likely that ice grounded on this shoal has acted to anchor fast ice in the prodelta region.

From the descriptions given above, it should not be assumed that fast ice around the perimeter is flat and featureless. We have already described an ice-piling event observed off Nome in 1973. Echert (personal communication, 1979) describes three rubble piles within the Norton Sound fast-ice zone: one approximately two miles east of Point Dexter, Norton Bay, $100 \times 33 \times 8$ m in size, at a water depth of 5 m; another approximately two miles west of Stuart Island, $150 \times 67 \times 8$ m in size, at a water depth of 5 m; and a third approximately eight miles north of the Apoon (north-flowing) mouth of the Yukon River, $230 \times 100 \times 5$ m in size, at a water depth of about 3 m. These rubble piles were well within the fast-ice zone and were probably formed during the initial freezing process.

Ice behavior within Norton Sound

The central part of the Norton Sound basin is

relatively flat, varying from 18 to 30 m in depth. Nelson et al. (this volume) report observing apparent ice keel gouges at water depths up to 22 m. Hence, at least occasionally, deep-draft ice features may be found within this basin. However, there is also evidence that a significant part of Norton Sound is covered by relatively thin ice not likely to pile up enough to attain such great keel depths: a large polynya in the northeastern sector is a virtually constant feature of Norton Sound. New ice is usually being formed here and transported westward by winds (see arrows indicating prevailing wind direction on Fig. 11-8) toward the entrance to the sound. Long before it arrives there, considerable thickening takes place through compaction and normal thermal accretion. The polynya is created by almost constant north-northeasterly winds across the northern shore of eastern Norton Sound and nearly constant easterly winds across the western end of the sound (Brower et al. 1977), as shown by the wind direction arrows on Fig. 11-8.

Figs. 11-9 and 11-10 are LANDSAT images of portions of Norton Sound, illustrating the range of ice conditions described above. Fig. 11-9, scene number 2399-21443 obtained on 25 February 1976, shows the entrance to Norton Sound. The Sledge Island polynya can be seen left of center. The Yukon Delta can be seen in the lower right corner with extensive fast ice on the Yukon prodelta. Several different ages of ice are apparent on the image. The ice within the entrance to Norton Sound and that far outside the entrance are both older than the band of ice extending down the Bering coast past the entrance to the sound. In the few days before this image was taken, the Bering Sea pack moved offshore approximately 20 km, allowing the band of newer ice to form. At the same time, older ice moved out of Norton Sound past the mouth of the Yukon and to the south. Hence, a band of large pans can be seen spilling out of Norton Sound to the south.

Fig. 11-10, scene number 2397-21330, was obtained two days before the scene just described. This scene shows inner Norton Sound; there is approximately 30 percent overlap with the area of the previous scene. Of interest here is the Norton Sound polynya, which opened up during the recent ice-moving event and has since frozen over with new ice. Nearer the entrance to the sound, there is young ice which was moved out of the polynya area during the event. This ice was probably new ice at that time. This image illustrates the cyclic nature of the Norton Sound polynya. It also shows the variability of the fast ice on the eastern side of the sound: a large chunk of attached fast ice has recently been de-

tached and has subsequently been broken into pans. The voids between the pans then froze. A very recent ice-breaking event resulted in a series of fractures running through this region on a southwest-northeast trend.

Pack-ice behavior along the western Alaskan coast—Bering Strait to the mouth of the Yukon

Various authors (McNutt, Nelson et al., and Pease, this volume) have noted that Bering Sea pack ice is generally moving from north to south past the west coast of Alaska. Although the winds at Nome have a significant easterly component, ice from the north of St. Lawrence Island has been observed to pass around the eastern side of the island and proceed along the western edge of the Yukon prodelta. Two distinct regimes of ice motion are often observed in this vicinity. Cox (personal communication) has prepared maps based on satellite imagery showing the Bering Sea pack-ice motion southward past the end of Norton Sound blending with the motion of ice outbound from Norton Sound along a line of shear at the entrance. A distinct line of demarcation can often be found running from north to south, dividing these two ice regimes. Ray (this volume) has also analyzed ice motions in this vicinity in detail and has concluded:

In general, the seasonal pack ice in Norton Sound is largely derived *in situ*, and tends to flow to the west and southwest in response to the prevailing winds or to flow sluggishly (eastward) in response to relatively weak oceanic currents during periods of relatively weak winds.

Along the western side of the Yukon prodelta is a region where the seasonal average ice edges nearly coincide. Their location agrees well with the edge of the prodelta where water depths change abruptly from 2 to 12 m. Occasionally grounded shear ridges have been observed along this zone, but their presence has not significantly increased the extent of fast ice. Although Bering Sea pack ice is often driven into this region, the edge of fast ice has not been observed to build out to the 20-m isobath as it does in the Beaufort Sea under somewhat similar conditions.

South of the Yukon Delta to Cape Romanzof, the edge of fast ice exhibits a great deal of variability. This is due at least in part to two shoals 60 km offshore at depths of 8 and 10 m and other shoals at intermediate distances. During the winter months, ice appears likely to pile on the outer shoals, forming anchors for extending the fast ice.

Just south of Cape Romanzof, the average fast-ice edges are again highly coincident. Then, approaching Nelson Island, they vary significantly with time.

Figure 11-9. LANDSAT image obtained 25 February 1976 showing the entrance to Norton Sound.

It is difficult to ascribe any particular cause to this behavior. A rather abrupt bathymetric transition from 6 to 16 m crosses this zone. The more closely coincident part of the average ice edge agrees well with this break, whereas farther south the ice edges oscillate across the break. The coincident part lies along the edge of mud flats which mark the prodelta of a former mouth of the Yukon River; the oscillating part lies across the mouth of Hazen Bay. Possibly the edge of fast ice to the north is determined by ice bottom-fast to the mud flats, and to the south the irregularity is caused by tidal currents into

and from Hazen Bay. The depth of the bay is between 4 and 5 m and the tidal range, which is as great as 3.5 m diurnally at Cape Romanzof, is sufficient to cause high-velocity tides into and out of the bay.

South of Nelson Island, around to Cape Avinof, the seasonal edges of fast ice coincide again. These boundaries appear to coincide with the 8-m isobath. Offshore from here, as far south as the southern side of Nunivak Island, there are several shoals 4-6 m deep as far as 30 km from the coast. Ice passing from north to south through Etolin Strait between the mainland and Nunivak Island often piles on these

Figure 11-10. LANDSAT image obtained 23 February 1976 showing central portion and eastern end of Norton Sound. Note large polynya at eastern end of sound.

shoals, creating relatively large (several km) islands of grounded ice.

Ice around Nunivak Island

Despite the large flux of ice down the Bering coast and the existence of several relatively shallow shoals (6 m) just to the north, Nunivak Island does not seem to retain an extensive expanse of fast ice. On the north side, facing the flow of ice southward along the Bering coast, fast ice bridges the embayments between Cape Etolin, the peninsula at the northern tip

of the island, and Cape Mohican at the western end, passing quite close to each headland and right along the bluff at Cape Mohican. To the east of Cape Etolin, fast ice extends farthest from shore and is found at its greatest depth in the waters near the island. Just to the east of Cape Etolin, the edge of fast ice in all periods comes close to the 20-m isobath. It is likely that ice passing through Etolin Strait is driven into this area, forming extensive expanses of fast ice. However, the period-average edges of fast ice on the east side of the island are all together well

inshore from the 10-m isobath and quite close to shore. Tidal flushing through Etolin Strait probably causes this behavior.

At Cape Corwin, the southeastern prominence of the island, the edge of fast ice is tangent to the curve of land forming the cape. From there, the fast ice extends across a wide bay to Cape Mendenhall, where it is again tangent to the coast.

From Cape Mendenhall to the southwestern edge of the island, the 20-m isobath is close to the edge of the island. It is unlikely that any fast ice is grounded here. The edge of fast ice is close to shore except for an area where it bridges a wide bay just to the west of Cape Mendenhall. This ice is not likely to be grounded, but it remains fast because it is protected from the general north-to-south ice motion past the island.

From Cape Avinof to the mouth of the Kuskokwim River, the edge of fast ice follows the edge of extensive mud flats on the north of Kuskokwim Bay. Although some variation can be seen, for the most part the fast-ice edge is consistently from year to year and period to period within each season on the finger-like projections of these mud flats. Individual LANDSAT images often show evidence of tidal flushing here. Large blocks of ice are broken loose and transported further offshore or set adrift. Further into the bay, there are several uncharted shoals where fast ice is frequently found.

Fig. 11-11 shows LANDSAT scene 1220-21440, taken on 28 February 1978. This scene illustrates the Etolin Strait region, with Nunivak Island on the left and the mountainous Nelson Island left of top center. The scene clearly shows the motion of ice along and away from the coast in this area. There is open water along most of the coast and on the south side of Nunivak Island. Fast ice can be seen on the shoals within Etolin Strait. The southward motion of ice through the strait is illustrated by the polynyas on the south side of these shoals. Farther offshore, older and thicker pack ice can be seen embedded in a matrix of newer ice, illustrating the continuous break-up of pack ice and formation of new ice in this region.

In the mouth of the Kuskokwim River is an embayment of the fast-ice edge which reaches a considerable distance upriver and is consistent in location from period to period. The Kuskokwim is navigable by oceangoing ships far past this point and has a reasonably deep channel. The tidal range here is 5 m with the diurnal range around 4 m. There is little doubt that the large tidal fluctuations are responsible for keeping this area free from fast ice.

Around the east side of Kuskokwim Bay to Jacksmith Bay, the edge of fast ice is on shoals 5-10

km offshore. Here the edge of fast ice moves farther from or closer to the shore with the changing seasons. From Jacksmith Bay to Goodnews Bay, the edge of fast ice follows the coast very closely despite the presence of extensive shoals further offshore at depths of 2-4 m. The diurnal tidal variation here is approximately 3 m—probably enough to remove any ice that might be temporarily grounded on these shoals. Winds in this vicinity come characteristically out of the northeastern quadrant and tend to remove ice from this coast rather than to cause ridge-building.

From Goodnews Bay to Cape Newenham, the edge of fast ice in winter and early spring (later there is none) bridges a wide embayment with depths on the order of 8 m. There is one shoal at a depth of just over 2 m north of Cape Newenham, but the ice does not appear to be anchored there.

From Cape Newenham to Naknek along the northern side of Bristol Bay, fast ice is found only in well-protected embayments at water depths generally less than 4 m. This is largely the combined effect of extreme tidal variations (7 m average diurnal range at Naknek) and strong offshore winds: (1) tidal variations break up and raft away ice not firmly anchored in place; (2) the pack ice is free to move towards the southwest; (3) the prevailing winds are toward the southwest.

These conditions generally prevail from Naknek to Egegik Bay. From about Egegik Bay southwestward there tends to be more fast ice. However, the presence of ice here depends in part on the severity of the ice year and in part on meteorological events required for ice to occur here. On the first of March in 1974, a year of heavy ice (Niebauer, this volume), winds drove the Bristol Bay pack ice onto the shore of the Alaska Peninsula, creating an extensive area of fast ice, including massive ridges. These ridges were some distance inshore of the 20-m isobath, however, and were probably formed from relatively thin ice.

Bristol Bay ice conditions

Pack ice in Bristol Bay appears to be greatly influenced by the fact that no barrier exists to keep ice from moving to the southwest. This circumstance, combined with the presence of strong offshore prevailing winds around the perimeter of the bay, results in a general southwestward motion of ice out of Bristol Bay. Normally, this motion is so persistent that LANDSAT and lower-resolution satellite imagery nearly always show open water along the northern side of the bay. As explained earlier, fast ice is not extensive and is generally found only in highly protected locations.

Figure 11-11. LANDSAT image obtained 28 February 1973 showing Nunivak Island, Etolin Strait, and the area of extensive mud flats west of the mouth of the Kuskokwim River. Note ice stranded on shoals in lower Etolin Strait.

Due to the nearly constant motion of ice away from the coast and the resulting open water, new ice is often forming along a broad band running east to west all across the northern side of the bay. It is often possible to clearly see the transition from open water to new ice, young ice, and first-year pack

ice on a single LANDSAT image. Superimposed on this behavioral pattern is a second characteristic: as the ice moves out of Bristol Bay into a less confined area, it breaks up into large pans with dimensions on the order of 10-20 km. The voids between these pans then freeze. This new sheet may then break up,

Figure 11-12. LANDSAT image obtained 9 March 1964 showing the central portion of the north side of Bristol Bay.

followed by the freezing of the new leads and voids. Evidence for several cycles of this activity can often be seen.

Fig. 11-12 shows LANDSAT scene 1594-21160, for 9 March 1974. This scene shows ice conditions along the northern side of Bristol Bay. Open water can be seen on the lee side of the land and adjacent

islands. Farther offshore, a stepwise gradation to thicker, older ice types can be seen, illustrating that the ice moves in accordance with a series of discrete ice-moving events. This scene illustrates why buildup of extensive fast ice in this region is rare, requiring unusual circumstances. Although the characteristic motion is out of Bristol Bay, occasionally a storm can

Figure 11-13. NOAA satellite image obtained 19 March 1975 showing entire Alaskan Bering Sea coast under study. The conditions seen here are typical and tend to support the description of Bering nearshore ice characteristics generated by this study.

drive ice onto the coast. One such event was mentioned in describing the behavior of fast ice along the Alaska Peninsula.

CONCLUSIONS

Nearshore ice conditions along the Alaskan Bering coast exhibit a wide range of characteristics from Cape Prince of Wales to the Alaska Peninsula. From Wales to Sledge Island, conditions are largely the same as those along the northern Chukchi Sea coast, while off Nome, fast-ice conditions are often similar to those of Beaufort Sea fast ice. However, further south along the coast, fast ice is found only in shallow and shielded areas. Finally, along the perimeter of Bristol Bay, fast ice is found only on mud flats and the upper reaches of estuaries.

Fig. 11-13 is a NOAA satellite image of the entire Bering coast obtained on 19 March 1975. This image illustrates the relationship between pack ice and fast ice described previously.

Fig. 11-8 was compiled to present in one figure the general characteristics of Bering Sea nearshore ice. Although there is considerable agreement between the satellite image in Fig. 11-13 and Fig. 11-8, it should be borne in mind that Fig. 11-13 is a selected, instantaneous satellite image and that the characteristics of Fig. 11-8 are based on an *average* of ice statistics. Although the schematic diagram presented in Fig. 11-8 is generally correct, it should be stressed that specific conditions produce variations from the average.

Beyond Fig. 11-8, there are four factors responsible for the characteristic behavior of Bering Sea ice:

1. There is a general moderation of climate as one moves southward, resulting in conditions limiting the growth of ice.
2. There is a prevailing motion of pack ice toward the ice front in the southern Bering Sea. As a result, coastal pack-ice movements are either directly away from the coast or along the coast. This motion causes large polynyas to open up in many areas.
3. The prevailing winds are directly offshore from all south- and west-facing coasts. As a result, newly formed ice in the polynyas is transported away from shore.
4. From Nome to King Salmon, the diurnal tidal range increases from 0.5 to 6 m. These tidal variations tend to lift ice away from the sea bottom so that it may be transported by currents and winds. One source of strong seaward currents is the large tidal variation. These currents can be particularly strong in areas of mud flats with large expanses of shallow water.

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Bering Sea Ice-edge Phenomena

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ABSTRACT

Recent field work at the ice edge during March 1979 shows that the interaction of ocean swell and winds contributes to the nature of the ice edge. Our observations show that the edge divides into three zones. At the seaward edge, there is a zone approximately 10 km wide which consists of heavily rafted and ridged ice floes with heights on the order of 1 m, depths on the order of 5 m, and diameters of 10-20 m. The second zone, which is approximately 5 km wide, consists of rectangular ice floes about 20 m across and 0.5 m thick which have been broken by the ocean swell, but are not heavily rafted or ridged. In the third zone, the ocean swell propagates through the ice without fracturing it, so that the floes have horizontal length scales of km, and thicknesses of about 0.3 m. We also found that when the wind blew strongly off the ice, the outer zone, because of its increased aerodynamic drag, moved downwind ahead of the rest of the pack to form the characteristic bands observed by satellite. These bands, which are on the order of 1 km wide and 10 km long, continued to move south into warmer water until they disintegrated under the action of warm water and waves.

INTRODUCTION

In this chapter, we review the ice-edge properties which we observed during our March 1979 OCSEAP *Surveyor* cruise. In the first part, we show that the ice adjacent to open water divides into three zones, which we call the edge, transition, and interior zones, following Squire and Moore (1980). These zones form because of the interaction of both ocean swell and wind stress with the pack ice. In the second part of the paper, we show that the increased aerodynamic roughness of the edge zone leads to the formation of long, linear bands of ice, measuring about 10 km in length and 1 km in width. These bands form nearly perpendicular to the wind during periods of off-ice winds and move southwest ahead of the pack ice. We document from observations of the formation and movement of a single band the fact that the band moved about 30 km southwest of the pack into warmer water.

THE NATURE OF THE ICE EDGE

Fig. 12-1 shows the location of the ice cores taken during the cruise, where the southernmost stations show the location of the ice edge. We took the ice cores using a helicopter operating from the *Surveyor*. In this section, we discuss the ice-edge properties based on the traverse lines labeled W, B, and C, which were respectively occupied on 6, 7, and 9 March. In the next section we first review the general properties of the ice edge, then document the specific ice properties observed on the three traverse lines.

Fig. 12-2 shows a schematic diagram of the three kinds of ice, in both plan and side view. First, at the outer edge of the pack, there is open water. Then the edge zone, 5-10 km wide, consists of small broken floes which are about 10-20 m in diameter. In cross section, these floes are heavily rafted and ridged, with sail heights of up to 1 m, and keel depths of 2-4 m. Figs. 12-3a and 12-3b show from photographs taken on 9 March an aerial and surface view of the ice near the edge. The ridge in the foreground of Fig. 12-3b, which is 1 m high and made up of ice 0.1-0.2 m thick, is shown from the air in the center foreground of Fig. 12-3a. The reason for the heavy rafting and ridging in the edge zone is that both ocean swell and wind act on the floes to work them against one another. Squire and Moore (1980) show from data taken on the same cruise that the swell propagates at least 65 km into the pack. From their data, the observed predominant wave period was 8-10 sec, which yields a wavelength of 100-160 m, so that the 20 m floe diameter is a fraction of a wavelength.

The second, or transition zone, about 5 km wide, is characterized by a tiled checkerboard pattern.

Figure 12-1. Chart showing the location of all ice cores taken during the March 1979 Surveyor cruise (chart courtesy Carol H. Pease).

Figure 12-2. A schematic diagram of the three kinds of ice which occur near the ice edge, proceeding inward from open water. The upper part of the figure shows the ice in plan view; the lower part, in side view.

Figure 12-3. Floes in the edge zone. (a) Aerial view; (b) surface photograph of the ridge in the foreground of (a). See text for further explanation.

Figs. 12-7a, 12-11b, and 12-14 show the most dramatic examples of this patterned ice, which consists of rectangles measuring about 20 m in width and 40 m in length, with their long axes perpendicular to the direction of wave propagation. In cross section, this ice is about 0.3-0.6 m thick; the thicker floes consist of two or three rafted pieces. This zone exists because the outer ice zone reduces the swell amplitude to the point that the waves fracture the ice without heavily rafting or ridging it.

At the inside edge of the transition zone, we observed an abrupt transition from the patterned ice to large floes measuring kilometers in extent. This third interior zone is the region of large floes where the swell amplitude is reduced so much that it propagates elastically without fracturing the ice. Fig. 12-8a shows an aerial view of this ice, where the floes measure kilometers in extent and 0.2-0.3 m in thickness. Satellite images suggest that this zone extends far to the north. In support of this general ice-edge picture, we next examine the specific properties along the three traverse lines.

Line W

The W-traverse took place on 6 March at the positions shown on Fig. 12-1. The wind on this day was negligible and the air temperature was about -5°C . Fig. 12-4 shows a schematic diagram of the kinds of ice observed on the traverse; the ice consisted of an edge zone 8 km wide and a transition zone 3 km wide with an abrupt transition from the rectangular floes to the interior zone at the northern edge of the transition zone.

The outermost floe on this line is station Scott. We occupied this floe for a strain-gauge experiment in the morning of 6 March; we were also fortunate enough to have divers from the ship, namely Lt. Comdr. Turnbull, Lt. Williscroft, Lt. (jg.) Fox, and Mr. Kramer, survey the bottom ice profile. Fig. 12-5a shows an oblique aerial photograph of station Scott from an altitude of 60 m after the strain-gauge experiment and before the diving. The dark area toward the camera is called the beach, and a small ridge-crack system runs horizontally in the photograph across the floe. The floe measured about 23 m by 37 m. Fig. 12-5b shows a surface photograph of the floe sighting down the ridge-crack system toward the beach; other ridged floes are visible in the background. We deliberately chose this floe because it was flat, unlike some of the surrounding floes which were heavily ridged.

Even though the floe surface appeared flat, the under-ice topography was very rough. To illustrate,

Figure 12-4. A schematic diagram of the kinds of ice observed along Line W. The circles indicate edge-zone ice; the small squares, transition zone ice. Stations Scott, W1, and W2 show coring site locations.

Fig. 12-6a is a map of the surface made by Vernon Squire, showing the location of the survey lines DEC and BEA; the dashed line shows the approximate place where the floe fractured before the diving. Fig. 12-6b shows the ice thickness from the two under-ice traverses. The divers gathered this information by stretching lines with knots at 10-ft intervals beneath the ice, then recording the depth at each knot from their wrist pressure gauges. Their observations showed that the bottom was very rough, with a maximum depth of 3.5 m.

Fig. 12-7a, from the transition zone, shows an aerial photograph of the floes at station W1. These floes measured 10-25 m on a side; the one on which we landed was 0.66 m thick and showed evidence of rafting at depths of 0.23 and 0.51 m. Fig. 12-7b shows a surface view from our floe; the pressure ridges behind the helicopter were about 1 m high. When we landed on the floe, we could feel the ocean swell propagating through the ice and see the swell-

Figure 12-5. Floe Scott. (a) An oblique aerial photograph from 60 m. Dark area toward camera is called the beach; small ridge-crack line runs horizontally across the floe. (b) Surface view sighting down the ridge-crack line towards the beach.

Figure 12-6. The shape of floe Scott. (a) Surface map of Scott (courtesy Vernon Squire); (b) two under-ice cross sections of Scott. Letters A, B, C, D, E on cross sections refer to survey lines on (a).

induced relative motion of the surrounding floes. We also observed at least seven walrus on the ice within 100 m of our station, which is further evidence (see Burns, *Ice As a Marine Mammal Habitat*, Volume 2 of this book) that these large floes surrounded by open water serve as habitat for walrus.

Fig. 12-8a is an aerial photograph from an altitude of 150 m of site W2 in the interior zone, where we landed on the large floe in the center of the photograph. The ice further in from this station also resembled this floe; the floes were large and flat, with a low-amplitude swell propagating through them. Fig. 12-8b shows the surface view; the ice was 0.24 m thick and covered by about 5 mm of snow.

Line B

The B-traverse took place on the next day, 7 March, at the positions shown on Fig. 12-1. As on the previous day, the wind was negligible and the air temperature was about -5°C . Fig. 12-9 is a sketch of the ice properties along the traverse line; again, the ice consisted of an edge zone measuring about 13 km wide and a transition zone 3 km wide with an abrupt transition from the rectangular floes to the large interior floes.

For the stations along the traverse, namely the ship, B4, and B5, Fig. 12-10a first shows an aerial view of the 90-m-long ship. In the center foreground, a party working on the ice is visible; Fig. 12-10b shows a close-up of this party from the ship. Again, the floes measured 10-20 m across and coring observations showed that the occupied floe was more than 1 m thick. Many of the floes seen near the ship had wetted surfaces, caused when the swell washes water onto the ice surface.

Further into the pack, Fig. 12-11a and 12-11b, taken from an altitude of 300 m, show the ice in the edge and transition zones. In Fig. 12-11b, the rectangular broken ice pattern is prominent, and the ice floes are wetted along the recently broken edges. For station B5 in the edge zone, Fig. 12-12a shows an aerial view of the floe on which we landed. This floe was about 20 m across and 0.34 m thick. Fig. 12-12b shows the surface appearance of the surrounding floes, and Fig. 12-12c shows the wetted edge of the floe, with a pen on the ice for scale. Finally, at station B4 in the interior zone, the ice again consisted of large flat floes 0.26 m thick. This traverse line again shows that the smallest, thickest floes occur at the ice edge.

Figure 12-7. Site W1. (a) Aerial photograph from 75 m of nearby floes; (b) surface photograph.

Figure 12-8. Site W2. (a) Aerial photograph from 150 m; (b) surface photograph.

Figure 12-9. A schematic diagram of the kinds of ice observed along Line B. Arrows mark the coring stations; see also legend for Fig. 12-4.

Line C

The C-traverse took place on 9 March 1979. On 8 March the weather deteriorated, consisting of blowing snow from the northeast with air temperatures of -1 to -3 C. On 9 March, the wind was 5.5 m/sec from the northeast at temperatures between -4 and -5 C. The ice was beginning to rot, and there was a strong southerly swell propagating into the pack. Fig. 12-13 shows the traverse line. Because of the wind, the ice was more open along the line, and the smooth progression from small floes to large floes was not evident as we flew north. Rather, the ice consisted of bands of open water and intermingled large and small floes for the first 30 km, at which point we reached the interior zone. We landed at two stations on the traverse line, C4 and C3.

C4 was a large floe consisting of ice which was mushy except in the bottom few centimeters, 0.16 m thick with a minimum ice temperature of -3.1 C. On

the floe surface, large amplitude 8 sec waves propagated through the ice on which we stood. By contrast, station C3 was on a large floe immediately adjacent to the rectangular broken ice; Fig. 12-14 shows an aerial view from an altitude of 150 m. This picture shows the swell breaking off rectangular floes about 20 m wide from a much larger floe; again the long axis of the rectangular floes was at right angles to the direction of wave propagation. We landed on the large floe shown in the left foreground of Fig. 12-14 to observe that the ice was firm and 0.33 m thick, with a minimum ice temperature of -4.3 C. Therefore, the floe distribution shown in Fig. 12-13 is caused by the fact that when the thin floes are warm, they are more elastic, so that the swell propagates through them without fracturing them; whereas the colder thicker floes, as shown in Fig. 12-14, fracture as the swell passes through them.

THE FORMATION AND MOVEMENT OF THE ICE-EDGE BANDS

The most interesting effect of the ice distribution described in the previous section is that it leads to the formation of ice-edge bands. Muench and Charnell (1977) review the satellite observations of the bands of ice which form at the edge of the Bering Sea pack ice during periods of off-ice winds. They show from analysis of satellite data that these bands have lengths on the order of 10 km and widths on the order of 1 km, and that the long axes of these bands are generally oriented at 40 - 90° to the left of the wind. They speculate that these bands form due to surface convergences caused by atmospheric roll vortices.

These bands also form at the edge of ice packs in other seas; Campbell et al. (1977, Fig. 5) document their formation in January 1974 in the Gulf of St. Lawrence from Skylab photographs; and this author has seen NOAA imagery of their formation at the edge of the Antarctic pack ice. Fig. 12-15, an aerial photograph taken from the NASA Convair-990 over the Bering Sea in 1973, Julian day 60, 00 h, 06 min, 30 s at 59.87° N, 175.063° W shows these bands (from Anonymous 1973). The aircraft heading was 002 - 004° and its altitude was 10.5 km with 70° field-of-view, so that the area depicted measures 14.7 km across.

Figure 12-10. The ice near the ship. (a) Aerial view; (b) surface view of ice party.

Figure 12-11. Aerial views from 300 m of the ice along Line B. (a) Edge zone; (b) transition zone.

Figure 12-12. Site B5. (a) Aerial view of floe, surface of which is marked by footprints. (b) Surface appearance of surrounding floes; (c) surface of floe B5.

Figure 12-13. A schematic diagram of the kinds of ice observed along Line C. The scale is half that of Figs. 12-4 and 12-9. The region marked 30 percent open water above Station C3 consisted of large floes with leads.

Figure 12-14. Site C3. Aerial view from 150 m floes broken by swell adjacent to large floes on which we landed. Helicopter antenna is at lower right.

In Fig. 12-15, beneath the cloud bands which are aligned approximately parallel to the wind, ice bands are visible with their long axes aligned at approximately right angles to the wind. The bands are 10-13 km long and about 0.6 km wide at their widest point. For the same day, Gloersen and LaViolette (1974) show the surface pressure map and a U.S. Air Force Weather Service Satellite image of the Bering Sea ice. The satellite image shows numerous ice bands at the edge; the weather map shows that the winds are approximately from the northeast. Coast Guard weather data gathered at the same time (from Campbell et al. 1975) show that the surface air temperature was -10°C and the wind speed was 5 m/sec.

How do these bands form? In the previous section, we showed that the effect of wind and waves at the ice edge was to work the ice in the edge zone so that it consists of numerous small floes, heavily rafted and ridged. Therefore the pack ice has an outer band of much thicker ice, with considerable top and bottom topography, adjacent to the open water. We therefore carried out from the *Surveyor* a field experiment during a period of off-ice winds to study the movement of this thicker edge ice relative to the thinner

ice in the transition zone. We found that because of the increased aerodynamic roughness of the ice in the edge zone it moved faster downwind than the interior ice. This relative velocity increase leads to the formation of the observed bands.

The experiment

On the morning of March 11, after a day of weak easterly winds which compacted the ice edge, we used the helicopter to place six targets on ice floes which were then tracked over a 23-hour period and their position determined from the Global Navigation System (GNS-500A) on the helicopter. Each target consisted of a colored nylon rectangle about $1 \times 2 \text{ m}^2$, nailed to the ice. Also, on three of the floes, we set up wooden poles 2 m high with radar reflectors mounted on top. Although we found that the reflector signal could not be distinguished from the ice on the radar of the ship, the poles served as good visual targets.

All of the floes tagged were within 3 km of the ice edge. The floes tagged with yellow, purple-1, red,

Figure 12-15. Aerial photograph from 10.5 km of the ice edges bands; north is to the top. See text for further explanation (photograph courtesy NASA).

and green were snow-covered and rafted cakes measuring about $20 \times 15 \text{ m}^2$. Fig. 12-16 shows the appearance of the floe with the green target (the green floe) shortly after the target was placed. The floe had a small ridge running down the middle with a rafted, snow-free area to the left. The purple-2 floe also measured $20 \times 15 \text{ m}^2$, and had a 0.6 m ridge to one side. Finally, the blue floe, shown in Fig. 12-17, measured only $3 \times 5 \text{ m}^2$ and had slightly raised edges. Although we did not measure the bottom topography of these floes, when we drilled holes 0.4 m deep for the poles, we did not reach bottom. We assume from our discussion in Section 1 that the floe thicknesses were on the order of 1-5 m.

We initially placed the targets in a cross, with the yellow, purple-1, red, and purple-2 floes aligned along the bearing $140\text{--}320^\circ$ over a distance of 1.4 km. On the other arm, the green, red, and blue floes were aligned along $042\text{--}222^\circ$ over a distance of 1.2 km, as shown at the top of Fig. 12-18. We then overflew the targets at 4, 9, and 23 hours from the time of placement and recorded their position. We also determined the distance and bearing from each target relative to the red floe. We lost some of the floes; purple-1 was not seen at four or nine hours, and blue disappeared after four hours. The evolution of the

targeted floes was as follows: when the targets were first placed, the ice edge was compact, with only a few ice cakes in the open water to the south, probably because of the weak easterly winds on the day before the experiment. Fig. 12-19 is a photograph looking from the ice toward the open water at this time, where in the original photograph the radar target mounted on the blue floe is visible in the middle of the picture.

Table 12-1 shows the averaged wind data for the 23-hour period. At four hours, after the wind began to pick up from the north, the targeted floes began to move away from the edge, creating an area with a 25 percent ice concentration inside an outer band of 90 percent concentration, as shown in Fig. 12-20. As Fig. 12-18 shows, the outer ice band formed a hook outward from the pack ice. The yellow floe was in the 25 percent ice concentration region, while the other targets were in the inner high-concentration region.

At nine hours (twilight so that photographs were impossible) all of the targeted floes were inside a band of ice floes 1-2 km across, several hundred meters southwest of the main pack. Our final 23-hour survey was on the morning of the next day.

Figure 12-16. Aerial view of the green floe; panel measures $1 \times 2 \text{ m}$.

Figure 12-17. Aerial view of the blue floe; pole is approximately 2 m high.

Figure 12-18. A schematic diagram of the observed advance and relative positions of the targeted ice floes during the 23-hour observation period.

TABLE 12-1

Averaged wind data for the 23-hour period

Time Period	Average Speed (m/sec)	Average Bearing (true degrees)
0-4	4.9	021
4-9	7.5	000
9-23	10.5	006
0-23	8.8	008

At this time, we observed many bands of ice similar to those shown in Fig. 12-15. Figs. 12-21a and b show two aerial views of a band near the ship similar to the targeted band. The targets were in the band shown in Fig. 12-18, which was now about 15 km south of the ice edge. Fig. 12-22 shows a schematic drawing of the band and the target positions at 23 hours. The band was about 2 km wide at its widest part, and 8-10 km long. It was widest at the end toward the swell, and had a long, curving tail.

Fig. 12-23 shows the way the band looked toward the tail; and Fig. 12-24 shows the way the head looked, sighting against the direction of swell propagation. Because of the action of the swell and the wind waves, this part of the band consisted of many small fragments of ice, pancake and larger, mostly

Figure 12-19. Aerial view from 75 m looking toward the ice edge as it appeared just after the placement of the targets. The blue floe is barely visible in the middle of the photograph.

Figure 12-20. Aerial view from 150 m of the ice edge as it appeared four hours after the placement of the targets. The targeted ice floes were mostly in the higher concentration of floes shown in the foreground.

Figure 12-21. Aerial view of a band of ice floes as it appeared the morning of March 12. (a) the 90-m-long *Surveyor* next to the band; (b) a section of the same band which was about 3 km east of the ship.

submerged floes which were covered with small pieces of ice. Finally, Fig. 12-25 shows the appearance of the green floe at 23 hours. To summarize, during our observational period, the red floe moved from $58^{\circ}56.6'N$, $170^{\circ}4.7'W$ to $58^{\circ}41.9'N$, $170^{\circ}22.4'W$, so that it travelled 32 km in the direction 210° , or at approximately 25° to the right of the average wind, at an average speed of 0.38 m/sec, or at 4 percent of the mean wind speed from Table 12-1.

The two major forces acting on the band, which account for its relative motion, are aerodynamic wind drag on the rougher floes in the edge zone, and the force which results from absorption and reflection of wind waves and swell. Due to the shape and roughness factors, the thicker, rougher ice in the edge zone has higher drag coefficients than the thinner, smoother ice found inside the edge. For example, Banke et al. (1976) found through measurements that sharp-edged, rough pancakes have air-ice drag coefficients two to four times larger than smoother, thinner floes. From a force balance, this suggests that at the same wind speed the thicker, rougher ice will move slightly faster than the smooth ice.

Figure 12-22. A schematic diagram of the band of ice in which the targeted ice floes were found after 23 hours. The location of the floes is marked using the same notation as in Fig. 12-18. The dotted portion represents the edges of the band characterized by small fragments of ice, pancakes, and larger, mostly submerged floes.

Thus, for wind speeds less than 5 m/sec, like those during the first few hours of the experiment, the thicker ice might move 1-2 cm/sec faster than the thinner ice. At wind speeds greater than 10 m/sec,

this value may increase to a 10-20 cm/sec difference in speeds. These numbers confirm what we observed: before the experiment the winds were low and easterly, so that due to the Coriolis force the ice motion would have been to the northwest. In addition, momentum transferred from the swell to the ice by reflection and absorption would have helped to form the compact ice edge which we observed at the start of the experiment. Then during the first four hours the winds, as Table 12-1 shows, came from the north as described above, and the targeted floes moved slightly south of the main edge. A speed difference of 1 cm/sec over four hours would cause the faster-moving ice to move about 150 m ahead of the slower ice. During the next five hours, the winds remained northerly and increased in speed, thereby increasing the difference in speeds and the distance between types of ice. In the final fourteen hours, the winds remained near 10 m/sec and the targeted floes were found some 15 km south of the ice edge. This requires an average difference in speeds of about 25 cm/sec. This number appears to be quite high when only wind and water stresses are considered; we must also consider the transference of energy and momentum from waves to the ice.

Once the bands move away from the immediate ice edge, the absorption and reflection of wind waves also exert a force on the ice. Longuet-Higgins (1977) shows that the force per-unit-width exerted on a floating mat by waves was equal to $\frac{1}{2} \rho g (a^2 + a'^2 - b^2)$, where ρ is the water density, g the gravitational acceleration, a the incident wave amplitude, and a' and b the reflected and transmitted amplitudes respectively. Wadhams (1973), through experiments with petri dishes in a wave tank, shows that the energy of short-period waves is almost all reflected or absorbed while that of longer-period waves is mostly transmitted with an exponential energy decay with distance into the ice. Dean and Harleman (1966) also suggest that for stationary and moored obstacles, the reflection coefficients are generally quite large when the horizontal dimensions of the obstacle are larger than half the incident wavelength. Thus, the reflected short-period wind waves enhance the motion of the band away from the main ice edge and also help to compact the ice into a band, because this force is primarily applied to the windward edge of the ice. Similarly, the enlarged head of a band may be the result of the absorption of swell energy at the leading edge of the band. These wave effects can enhance the ice speed by as much as 10 cm/sec, which brings the speed difference roughly to the desired 25 cm/sec. Thus, wave effects will contribute to the ice motion in the vicinity of the ice edge.

Figure 12-23. An aerial view from 75 m of the tail of the band in Fig. 12-22.

Figure 12-24. A view from 25 m of the small fragments, pancakes, and submerged floes at an edge of the band being acted upon by wind waves and swell and represented by dotted line in Fig. 12-22.

Figure 12-25. An aerial view from 50 m of the green floe as it appeared 23 hours after the target placement.

Consequences of the band motion

Table 12-2, which lists the air and water properties measured near the band during the period of our observations, shows that as the band moved south, it moved into warmer, more saline water. At the beginning of the experiment the water temperature was -1.2°C , and the salinity was $31.9^{\circ}/\text{oo}$. At the end of the 23 hours, the ice was in water of -0.45°C and a salinity of $32.2^{\circ}/\text{oo}$. In both cases, the freezing temperature of the water was about -1.7°C . The water temperature ranged from $+0.5$ to $+1.3$ degrees above freezing and we observed the ice to melt.

TABLE 12-2

Water and air properties observed from the ship near the band

Elapsed Time (h)	Air Temperature ($^{\circ}\text{C}$)	Water Temperature ($^{\circ}\text{C}$)	Salinity $^{\circ}/\text{oo}$	Freezing Point ($^{\circ}\text{C}$)
0	-0.70	-1.20	31.94	-1.73
4	-1.09	-1.30	32.05	-1.73
9	-1.79	-0.62	32.11	-1.73
23	-4.50	-0.45	32.23	-1.74
Average	-3.14	-0.60	32.11	-1.73

Since the band was relatively long and thin, lateral melting may have been of some importance, but vertical melting was most likely dominant because of the larger exposed surface area. Another form of melting is wave-induced; the waves fracture the ice floes, thereby creating for each floe a larger ratio of surface area to volume. Waves breaking over the ice can also cause melting of the upper surface. Wave-induced melting was visible along the windward and swellward edges of the band (Fig. 12-24). In summary, our observations suggest that these long thin bands of small floes rapidly melt in the warmer water.

From satellite images one can perceive the importance of this melting regime on a larger scale. In consecutive images, interior ice features are seen to move to the south or southwest throughout most of the eastern Bering Sea in late winter and early spring, as reported by Muench and Ahlnäs (1976) and by McNutt in this volume. However, the fact that the southern edge does not appear to advance south during much of this time suggests that the bands carry away enough ice to the southwest, where the ice melts, to maintain the ice-edge position. If this is indeed what is occurring, then a very large amount of ice is melted near the southern ice edge during periods of strong northeast winds.

CONCLUSIONS

From analysis of satellite data, several investigators, e.g., Muench and Ahlnäs (1976), McNutt in this volume, and Loshchilov (1973) conclude that much of the ice in the Bering Sea forms in the north, then is conveyed to the southwest by the strong northeast winds which accompany the anticyclonic circulation of the Siberian high-pressure system. The present investigation of the southern ice-edge processes shows in part how the ice-edge position is maintained.

The large ice floes of the pack ice are advected southwest by the wind. As they approach the ice edge, first the ocean swell propagating into the pack fractures them into the characteristic rectangular pattern; then in the edge zone the floes are heavily rafted and ridged, which leads to a great increase in aerodynamic roughness. Because of this increase in roughness the ice in the edge zone moves away from the pack with a greater relative velocity, and distributes itself into the characteristic bands. These bands continue to move southwest into warmer water until they disintegrate. Further, the southward movement and disintegration of the bands exposes the transition zone to higher swell amplitudes, so that this zone becomes the edge zone and in turn becomes heavily rafted and ridged, and blows southwest. In our observations, a band moved 32 km in 23 hours at an average speed of 0.38 m/sec into water that was 1.3 C warmer than its freezing temperature. Thus the formation and movement of the ice bands are one way in which the pack ice is rapidly dispersed and melts.

Moreover, these bands may play an important role in the dispersal of pollutants. First, oil spilled to the north which is trapped in the ice may enter this conveyor belt and be released 30-50 km south of the pack as an ice band disintegrates. Second, the bands generally had sharp leading edges and diffuse trailing edges, which suggests that the band moves faster than small pieces of ice or oil slicks. Therefore, for example, an oil slick southwest of one of these bands may be overtaken by the band and carried south to the point where the band disintegrates, so that the slick may spread over a larger area than would be anticipated from wind action on the slick alone.

DEDICATION

We dedicate this chapter to the memory of Robert Lewis Charnell, lost at sea near Hawaii in December 1978, who gave us help and encouragement during the planning phase of this ice-edge experiment.

ACKNOWLEDGMENTS

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Eastern Bering Sea Ice Dynamics and Thermodynamics

13

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ABSTRACT

During winter 1979, hydrographic, meteorological, and ice floe data were collected over the Bering Sea shelf. The ice pack extended to 59°N; however, there appeared to be little or no *in-situ* freezing in the study area. Hydrographic data from the marginal ice region (seaward limit of the ice) showed that less saline, cold (≈ -1.4 C) waters existed in an upper layer; the lower layer was as much as 1 C warmer. Floes advected toward the south to southwest at rates as high as 0.5 m/sec during north-to-northeast wind events. Floes rotted along the margin in periods on the order of days. Little ridging of ice was observed over the open shelf. Rafting was prevalent among floes battered by wind and swell at the ice edge.

We observed that in the fall, northerly winds cool the water of Norton Sound and the Bering Sea north of St. Lawrence Island until the water column is isothermal at freezing temperatures. Further cooling causes freezing. Under northerly wind conditions, ice is advected south into water where it is no longer in thermodynamic equilibrium. The resulting meltwater is mechanically mixed and is a source of cooling for the waters of the southern Bering shelf. These observations suggest that ice formation and movement in the Bering Sea can be likened to a conveyor belt: growth occurs primarily in the north, advection due to wind stress is generally southward, decay occurs at the thermodynamic limit, and the limit advances somewhat as meltwater cools the upper layer.

INTRODUCTION

On a cruise of the NOAA ship *Surveyor* during the first two weeks of March 1979, many types of oceanographic, meteorologic, and ice floe data were collected in order to identify processes inherent in the distribution and condition of sea ice over the Bering Sea shelf (Fig. 13-1). These data, in conjunction with simultaneous edge-specific studies by Martin and Bauer (Chapter 12, this volume), remote sensing studies by McNutt (Chapter 10, this volume), and previous work by Muench and Ahlnäs (1976) and Ahlnäs and Wendler (1979), yield a well-formed description of mesoscale interaction between dynamics and thermodynamics in controlling the pack ice conditions in the Bering Sea. This chapter describes and presents supporting evidence for this interaction.

Figure 13-1. Eastern Bering Sea ice study area including the observation area during the cruise of the NOAA ship *Surveyor* during March 1979.

ICE-EDGE HYDROGRAPHY

Forty-one CTD's (Fig. 13-2) and twenty-two airsondes (Fig. 13-3) were taken along the ice edge during the first two weeks of March 1979 from the NOAA ship *Surveyor*. Oceanic and atmospheric surface temperatures were taken every hour, and surface salinity was taken every two hours. The water temperatures were at first taken as bucket temperatures with a calibrated thermometer (± 0.2 C), but after the thermometer was lost, a calibrated

Figure 13-2. The location of CTD stations taken during the first two weeks of March 1979 from the NOAA ship *Surveyor*.

thermistor string (± 0.1 C) was used. This method also provided air temperature measurements near the surface. Although temperatures were measured within a few centimeters of the air-sea interface, they are considered representative of more extensive layers since turbulence from the ship's wake provided mechanical mixing. Salinity was determined from bucket samples using an auto-salinometer. With a motorized psychrometer, hourly dry-bulb and air temperatures were taken at the bridge throughout the cruise. However, problems were encountered with the wet-bulb thermometer readings and the data were discarded as unreliable. The bridge measurements were made approximately 10 m above the surface.

In general, the water temperature was above the freezing point of the ice except for that measured in CTD cast No. 1, 80 miles south of Nunivak Island in 30 m of water on March 2. Floes were rotting from the bottom west of this area. Melt puddles were not

formed as in the traditional summer Arctic melt condition, but some floes were wetted when waves washed over them. Details of the mechanical effects of the swell on the ice and resulting edge processes are discussed by Martin and Bauer (Chapter 12, this volume).

A major feature of the water column along the ice edge was fresh cold water due to ice melt. CTD sections perpendicular to the ice edge west of the Nunivak area showed a meltwater lens (Fig. 13-4). The pycnocline downwind of the pack edge was slightly deeper than that under the ice itself, apparently because ice cover protects the water column from wind-mixing; but the exact shape of the front is speculative. The ice in this figure was in surface waters of ≈ 1.0 C, although ice was observed in slightly warmer waters at other times.

We also observed the effect of melting by ex-

Figure 13-3. The location of radiosonde launches made during the first two weeks of March 1979 from the NOAA ship *Surveyor*.

aming the change in the distribution of sea surface temperature (Fig. 13-5). All surface measurements during the two-week cruise are plotted on the same chart and contoured for the first and second weeks separately. The eastern end of the figure shows the change observed over a two-week period, while the western end shows almost steady-state conditions. Note that while the -1.0 C isotherm moved about 35-40 km south during the cruise, the $+1.0$ C isotherm did not move appreciably. The salinity field showed the same trend; although the contouring is less significant with only half as many samples (Fig. 13-6), the advance of meltwater is strikingly apparent.

It is possible that movement of meltwater toward the south could be due to advection of the water mass itself. However, the mean currents on the shelf are weak and northwesterly, generally in opposition to the advance of meltwater (Kinder and

Schumacher, Chapter 5, this volume). Contours of dynamic height during the March cruise period also suggest weak flow toward the northwest (Fig. 13-7). Although this dynamic topography has less relief than that shown previously by Charnell et al. (1979), it is similar in inferred speed and direction. Also, the pattern suggested by Figs. 13-5, 13-6, and 13-7 is coherent over periods longer than a tidal cycle.

The amount of ice melt required to lower temperature to observed values (Fig. 13-4) was estimated by assuming that the water had been $+1.0$ C and that the ice had already warmed to its freezing point. Under these assumptions, enough heat was extracted to melt a 60-km-long, 0.5-m-thick strip of ice of 15‰ salinity. This melt estimate was probably too high since off-ice winds also cool the water, but it shows that the proposed physical mechanism is capable of producing the observed hydrographic features.

Figure 13-4. A CTD cross section perpendicular to the ice edge showing the meltwater lens typical along the margin of ice. The ice existed in water of about -1.0°C and colder. The CTD station numbers correspond to those in Fig. 13-2.

DISCUSSION

Ice forms *in situ* during a typical ice-year in late fall (November-December) in Norton Sound, in the

Bering Sea north of St. Lawrence Island, along the Alaskan coast, and eventually southward into Bristol Bay (Fleet Weather Facility 1972-1975, 1976, 1977, and 1978). These areas are shallow (less than 30 m), except for the region north of St. Lawrence Island, and water is observed to be isothermal at the freezing point (about -1.7°C). Cooling is caused by a negative radiation balance and sensible and latent heat transfer due to offshore winter winds. Under the influence of northerly to easterly winds, the ice floes are transported into warmer water. The leading floes melt, adding relatively cold, fresh water to the existing water column. Under continued northerly winds, the new leading floes encounter colder waters and thus have longer residence times. As the season progresses, regions of ice growth expand and the floes are continuously advected beyond these areas.

Satellite and aircraft photographs of the Bering Sea in winter show persistent polynyas in the lee of Cape Nome and St. Lawrence, St. Matthew, and Nunivak Islands during periods of northerly to northeasterly winds (Muench and Ahlnäs 1976, and McNutt, Chapter 10, this volume). Less commonly, a polynya can also be observed in eastern Norton Sound during more easterly wind events. Gray streaks, which Martin and Kauffman (1979) attribute to grease ice production, are often observed in satellite images of these polynyas. Floes 0.3-0.5 m thick near the ice edge but not yet rafted are formed of frazil in approximately the upper half, while the lower half grows in place as columnar crystals (Martin and Kauffman 1979). This indicates that floes continued to grow for a time after the grease ice had consolidated. Overflights show that the floes do not change much in thickness after they are advected away from shore and south of about 62°N until they are rafted and melted at the pack edge. Thus, we see areas of growth, relative thermal stability and transport, and thermal and mechanical destruction. These roughly correspond to the regions described by Muench and Ahlnäs (1976) in the spring of 1974. Because of these characteristics, the ice can advance 200-300 km beyond the fall growth region, while the growth region itself expands about 100-200 km beyond its fall extent.

Spring is heralded in the Bering Sea by shifting of the storm tracks north from the Gulf of Alaska to the southern Bering Sea (Overland, Chapter 2, this volume). Given sufficient time, the radiation balance alone would melt the ice pack, but the shifting of the winds from cold northerly to warm southerly can accelerate this procedure considerably. The entire pack can become uniformly rotten in three or four weeks after low-pressure systems penetrate the southern Bering Sea. A wind field during rotting

Figure 13-5. Hourly surface water temperatures between 2 March (JD-61) and 15 March (JD-74) 1979 from the NOAA ship *Surveyor*, contoured for the westward and eastward treks respectively. Note that the +1.0 C isotherm did not move significantly while the -1.0 C isotherm moved 35-40 km due to the inclusion of meltwater.

conditions occurred in late March 1979 (Fig. 13-8). The accelerated rotting was observed in winter 1974, a year with extensive ice cover (Muench and Ahlnäs 1976), and in 1979, a year with light ice cover (McNutt, Chapter 10, this volume).

The sea ice is generally divergent on the open Bering shelf because the wind field is slightly divergent when winds are northeasterly and the basin is less constrictive downwind. A northeasterly wind field occurred in early March 1979 (Fig. 13-9). The divergence resulted in little or no ridging of Bering Sea ice compared to the Arctic pack. The few observed ridges formed mainly windward of St. Lawrence Island and in southern Norton Sound. These features generally had low relief because there was little ice to build ridges. Parmenter and Coon (1973) showed that the maximum height of ridges is a function of ice strength and thickness of con-

tributing floes. Since the maximum ridge thickness is about 10 times that of contributing floes, we expect 3-5 m ridges from floes 0.3-0.5 m thick around St. Lawrence Island and possibly 10-20 m ridges from floes 1.0-2.0 m thick in southern Norton Sound.

Vertical water-column structure over the eastern Bering Shelf is complicated. It appears that in areas where the shelf is less than 50 m deep, the entire water column is tidally mixed. In water deeper than 50 m, the structure is two-layered with a pycnocline generally at about 20 m. The upper layer is mixed by wind and the lower layer by tides. Kinder and Schumacher discuss this in detail (Chapter 4, this volume). In ice-growth regions, vertical mixing is enhanced by the extrusion of salt during the freezing process. Since typical ice salinities are 10‰ or greater (Martin and Kauffman 1979), about 20‰ is extruded. A typical advection rate for floes is about

Figure 13-6. Bihourly ocean surface salinities between 2 March (JD-61) and 15 March (JD-74) 1979 from the NOAA ship *Surveyor*, contoured for the westward and eastward treks respectively. Note that the lighter isohalines have moved south during the two-week period while the heavier isohalines have not appreciably moved.

0.25 m/sec (McNutt, Chapter 10, this volume, Muench and Ahl $\ddot{a}$ s 1976), and we assume steady-state production versus advection of ice 0.3 m thick with a moderately persistent northerly wind regime for about four months. This is equivalent to growing over 5 m of ice in the growth region during the season. Growing this much $10^{\circ}/\text{oo}$ ice would increase the salinity of 20 m of water by about $5\text{--}6^{\circ}/\text{oo}$ over the season. With different choices for seasonal extent, velocities, and persistence, the range of salinity increase over the season is from $1.25^{\circ}/\text{oo}$ to $7.5^{\circ}/\text{oo}$. However, since the mean current is from the melt zone toward the growth zone, one would expect some of this salt to be recycled, thereby partially masking the salinity-enhancing process. The lower limit of these values is in agreement with those observed by Coachman et al. (1978) for the shoal region southwest of Nunivak Island.

A similar calculation can be made of the number of times the ice pack replaces itself. At a floe speed of 0.25 m/sec, assuming the pack extends 300 km south of Nome and persists for four months, the pack replaces itself about eight times in the season. With various choices for seasonal extent, velocities, and persistence, the range of possible replacements is from two to ten. It should be remembered, however, that the cycling of the pack is continuous. The number of replacements is an indication of the dynamic and thermodynamic vigor of the system.

SUMMARY AND CONCLUSIONS

During the winter of 1979, ice was observed to form in the northern and coastal regions of the Bering Sea, to advect south-southwesterly, and to melt along

Figure 13-7. Dynamic topography between 2 March and 15 March 1979 computed from CTD stations illustrated in Fig. 13-2. Note northwesterly set of lines of constant topography. This implies a relatively weak northwesterly current.

the southern margin; ice processes over the Bering Sea shelf can thus be thought of as a conveyor belt. The interaction between the dynamic and thermodynamic processes of decay along the margin was explored for March 1979 using measurements taken from the NOAA ship *Surveyor*. This analysis and remote sensing analysis by McNutt (Chapter 10, this volume) show that the same processes were at work during 1979, an extremely light ice year, as in 1974, a medium-heavy ice year (Muench and Ahlnäs 1976). This suggests that the extent of the pack ice from one year to another is a function of both the advective scheme and thermodynamic processes.

The conveyor belt process of ice growth and decay is responsible for enhancing salt content in northern and coastal waters (ice production areas) and reducing salinity, resulting in stratified waters along the ice

margin. This process has a considerable effect on the structure of water properties of the shelf, at least during the winter season. The vigor of the advective scheme, i.e., wind blowing the ice, also affects the extent of freezing.

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Figure 13-8. Sea-level pressure and wind fields calculated from digitized NWS-Alaska region surface analysis charts for 28, 29, and 30 March 1979. This pattern is the dominant spring condition driving the ice northerly and warming the basin. Longest vectors represent 20 m/sec winds.

Figure 13-9. Sea-level pressure and wind fields calculated from digitized NWS-Alaska region surface analysis charts for 1, 2, and 3 March 1979. This pattern is the dominant winter condition driving the ice southwesterly and cooling the basin. Longest vectors represent 20 m/sec winds.

ments were carried out by the survey technicians on the *Surveyor*, who patiently modified their operations to meet my requirements.

Frances Parmenter of the NESS field office in Anchorage, Bruce Webster, ice forecaster for NWS Fairbanks, and Doris Brown of the NWS Anchorage office contributed materials for the field parameter analysis. Lt. (jg.) Daniel V. Munger and AGC William B. Damico flew the Navy ice reconnaissance for the

NOAA-Navy Joint Ice Center under the leadership of Comdr. James C. Langemo.

Sigrid Salo of PMEL handled much of the data analysis and prepared the figures. Joy Golly and James Anderson completed the drafting and photography. Drs. James E. Overland and James D. Schumacher discussed important aspects of the physics. Seelye Martin and Lyn McNutt were my colleagues in the experiment.

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Anticipated Oil-Ice Interactions in the Bering Sea

14

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ABSTRACT

In the lee-shore regions of the Bering Sea, ice formation is dominated by the presence of wind waves and swell. Laboratory experiments and field observations show that the kinds of ice which form in these regions are grease and pancake ice. The laboratory experiments also show that much of the oil spilled within these kinds of ice accumulates on the surface. The chapter also summarizes the entrainment mechanisms for oil released under first-year ice, and discusses from field observations possible oil entrainment mechanisms which may occur at the ice edge.

INTRODUCTION

There are several ways in which oil and sea ice may interact in the Bering Sea. The interaction processes may be divided into those which capture oil and those which transport it. The capture processes include the interaction of oil with smooth, unbroken first-year ice, the entrainment of oil by rafted or ridged first-year ice and by grease and pancake ice, and the interaction of ocean swell, oil, and the ice floes at the ice edge. The transport processes include the interaction of oil with a Langmuir circulation in the lee-shore polynyas where grease ice occurs, the general advection of oil by large-scale ice movement, and the specific advection of oil by the ice bands which form at the ice edge.

OIL AND FIRST-YEAR ICE

Martin (1980) gives a detailed description of oil entrainment by first-year ice, based on both OCSEAP data and a field experiment carried out for the

Canadian Beaufort Sea Project. Fig. 14-1, which summarizes the evolution of a winter under-ice oil spill over the growing season, shows that the oil behavior divides into the following three parts:

November-February. When oil is first released under smooth first-year ice, it collects there in natural undulations, caused by snowdrift-induced insulation. These undulations have an amplitude of order 0.1 m and a length scale of order 10 m. Some oil also flows a short distance up into the ice through the brine drainage channels.

March-May. As the ice warms in the spring, top-to-bottom brine channels open up in the ice and the oil rises through these channels to the ice surface, where it spreads laterally, both under the snow and within the first few porous centimeters of the ice surface. The oil on the surface then absorbs solar radiation through the snow, causing the snow above the oil to melt.

June-August. In the early part of the summer, the trapped oil continues to rise through the ice to the surface, where above the trapped oil melt ponds with oily surfaces form. Because the sun heats the oil and the winds blow the oil on the melt-pond surfaces against the edges of the ponds, the oiled melt ponds grow both laterally and in depth faster than the un-oiled ponds. As the ice continues to decay throughout the summer, much of this oil in a weathered form will flow back into the ocean either off the tops of the floes or by melting through the ice bottom.

Figure 14-1. The interaction of oil with first-year ice (a) November-February; (b) March-May; (c) June-August. See text for further description.

RESULTS OF THE BUZZARDS BAY OIL SPILL

The Buzzards Bay, Massachusetts, oil spill of 28 January 1977, described in Deslauriers et al. (1977), yields additional information on how oil responds to tidal currents under rafted and ridged ice. Fig. 14-2a-c shows schematic diagrams of how oil which was initially swept under the ice came to the surface. In Fig. 14-2a and b, the strong 0.5 m/sec tidal currents sweep the oil into the lee of newly formed rafted ice, from whence it flows up into a surface pond, replacing the denser sea water. Then (Fig. 14-2c) the tidal current sweeps away the unsheltered oil into other areas of rafted ice. Fig. 14-3 shows the appearance of an oil-filled pond in plan view based on coring observations; these ponds were filled to a depth of 100-150 mm with oil. Estimates from aerial

photographs suggested that such ponds contained as much as 30 percent of the spilled oil.

Observations from the Buzzards Bay spill also show that oil trapped either on the surface or within floes will be transported a considerable distance by winds and currents. Fig. 14-4 shows an aerial view of an oiled ice floe measuring 75 m across moving through the 12-km-long Cape Cod Canal. The dark material on the floe surface is oil, and oil streaks are also visible on the open water. Observations showed that the floe moved through the canal into Massachusetts Bay, where the oil was released when the floe melted. This observation suggests that in the Bering Sea trapped oil may be carried distances on the order of hundreds of kilometers before the oil is released when the floe melts.

Figure 14-2. Interaction of oil in a tidal current with rafted ice. (a) Oil enters lee of raft; (b) oil fills surface pond; (c) current sweeps unsheltered oil away (adapted from Figure 2.3 of Deslauriers et al. 1977).

Figure 14-3. Plan view drawn from field observations of the oil pond shown in side view in Fig. 14-2 (adapted from Figure 2.2 of Deslauriers et al. 1977).

Figure 14-4. Aerial view of oiled ice floe in the Cape Cod Canal, 17 February. The highway next to the canal gives scale (from Figure 2.11, Deslauriers et al. 1977).

OIL IN GREASE ICE

Grease ice is a slurry of small ice crystals and seawater which grows in the marginal seas. Martin et al. (1978) give a preliminary discussion of how this ice grows, and Martin and Kauffman (1980) discuss exhaustively its general properties.

Grease ice in the Bering Sea

Fig. 14-5 shows a satellite image of the Bering and Chukchi Seas for 17 March 1976. The weather charts for this day and the preceding three days show that the winds were dominated by the Siberian high-pressure system, which created northeast winds over the Bering Sea. On 17 March 1976, the air temperature recorded at 1800 Z at Wales was -25°C and the wind velocity was 15 m/sec. The NOAA-4 image shows that one effect of these cold winds was to create lee-shore regions of dark, new ice formation where the older ice was blown away. These areas occur in Norton Sound from Wales to Norton Bay, on the south side of both St. Lawrence and Nunivak islands, in Kotzebue Sound from Point Hope to Kivalena, and along the Siberian coast.

For the same day, Fig. 14-6 is a LANDSAT image of the polynya south of St. Lawrence Island. The LANDSAT image, which covers an area 185 km across and has a 70 m resolution, shows that long linear ice plumes form approximately parallel to the

wind and extend 10-40 km downwind of the island, and that there appears to be a pile-up of new gray ice downwind of the ice plumes.

A close-up of these plumes is shown in Fig. 14-7, an aerial photograph of a polynya taken at a different time from the previous two images by the NASA Convair-990 on 20 February 1973 in the Bering Sea as part of the joint U.S.-U.S.S.R. Bering Sea Experiment. The aircraft flew at a radar altimeter height of 147 m and the camera had a 70° field-of-view, so that the field covered by the image measures 205 m across. The flight heading was 185° , and the recorded wind speed was 11.5 m/sec from 332° or from the upper left of the photograph, so that the plumes are again lined up parallel to the wind direction. The air temperature at altitude was about -14°C . From inspection of the photograph the predominant wavelength was about 1.5 m.

This photograph shows several interesting features. First, the grease-ice plumes have several different scales, ranging in width from less than 1 m to 35 m across (the plume marked *a* at the lower left). Second, the spacing between the plumes ranges from 2 to 4 m for the smallest bands to about 170 m between the wide plumes marked *a* and *b*; hence the size and spacing of the large plumes is such that they are visible on the LANDSAT. Third, there is an obvious pile-up of grease ice at the downward end of the plumes marked *c*, *d*, and *e*, accompanied by a

Figure 14-5. Composite NOAA-4 visible satellite image of the Bering Sea for 17 March 1976.

broadening of the head of the band. It is also clear, on the original negative at least, that there are no waves behind the heads *c*, *d*, and *e*. Dunbar and Weeks (1975) call these plumes "tadpoles" and show that pancake ice forms in the heads. They also suggest that an oceanic Langmuir circulation leads to the formation of the grease-ice plumes.

Fig. 14-8 shows an oblique photograph from 150 m of a polynya 3 km wide of Nome on 5 March 1978. The air temperature was -20°C , the wind speed was 15 m/sec from 010° , and the predominant wavelength was 6 m. The photograph clearly shows the organization of the grease ice into plumes parallel

to the wind, and the pile-up of grease ice downwind against the floes. The arrow on the photograph marks the location of a detritus line, which we will discuss later.

The field and laboratory work of Martin and Kauffman (1980) shows that the grease ice is a slurry of small ice platelets which individually measure approximately 1 mm in diameter and $1\text{--}10\text{ }\mu\text{m}$ in thickness. Fig. 14-9 shows a photograph taken through crossed polaroids of the clusters of ice platelets; the clusters which formed in our laboratory grease ice measure 1-5 mm across. Our observations show that the concentration of such crystals

Figure 14-6. LANDSAT image of the polynya south of St. Lawrence Island for 17 March 1976.

within the grease-ice plumes ranges from 15 to 45 percent and that the thickness of the ice varies from 0.1 to 0.3 m in the laboratory to more than 1 m in the field. Also, our experiments show that because grease ice has a nonlinear viscosity, which increases as both the ice concentration increases and the shear rate decreases, it is an excellent wave absorber.

Fig. 14-10, adapted from Pollard (1977) shows a schematic diagram of the Langmuir circulation which accompanies the formation of grease ice. The combination of the bi-directional wind-wave spectra and the mean wind drift leads to the formation of alternating roll vortices in the ocean, where the grease ice accu-

mulates in the convergence zones. This also suggests that, if oil is emulsified into small droplets by the wave breaking, these droplets may circulate around in the Langmuir rotors, as well as pile up in the convergence zones. Pollard (1977) states that typical convergence zone separations range from 5 to 300 m, and that rotor downwelling velocities are observed to be between 20 and 60 mm/sec. Further, the recent theoretical work of Craik and Leibovich (1976) suggests that the accumulation of a viscous substance such as oil or grease ice in the convergence zones will, by transforming the wave momentum into an additional mean current, intensify the rotor veloci-

Figure 14-7. Aerial photograph from 147 m of grease ice formation in the Bering Sea. See text for further explanation. Photograph courtesy NASA.

ties. This intensification could lead to the distribution of more oil throughout the water column.

The laboratory results

Our laboratory tank, described in Martin et al. (1978) and shown in Fig. 14-11, measured 2 m long,

1 m wide, and 0.6 m deep. In this tank, which was filled to a depth of 0.4 m with a 34‰ NaCl solution and placed in a cold room, we grew grease ice by generating waves. Fig. 14-11 shows the appearance of the wave decay and the pile-up of grease ice in the tank.

Figure 14-8. Oblique aerial view from 150 m of grease ice herded into Langmuir streaks south of Nome on 5 March 1978. See text for further description.

Figure 14-9. Photograph of clusters of frazil ice crystals taken between crossed-polaroids. The subject measures 21 mm in the vertical.

Figure 14-10. Schematic drawing of grease ice and the Langmuir circulation adapted from Pollard (1977, Figure 8.3).

Figure 14-11. Schematic side-view drawing of the experimental tank, with inner dimensions of 2 m in length, 1 m in width, and 0.6 m in depth.

To illustrate both the induced circulation and the form that oil entrainment takes in waves propagating into grease ice, Fig. 14-12 shows a composite image of a sideview photograph of the grease ice above a sketch of the wave-decay-induced mean velocity field in the grease ice; the sketch also shows the final distribution of oil released on the surface ahead of the grease ice. In the photograph, the grease ice is white, its maximum thickness is about 0.2 m, the wave-period is 0.54 sec, and the wavelength is 0.46 m. The paddle is to the left in the photograph and a wave probe sticks into the water ahead of the paddle; the gap in the photograph is caused by a tank support. Although we did not release oil into the grease ice shown in the photograph, the sketch shows the oil distribution which occurred in a similar experiment described below.

For oil entrainment, the most important property of grease ice is that it changes abruptly from liquid to solid behavior; the transition is shown in Fig. 14-12 as the "dead zone." Upstream of the dead zone, the

grease ice behaves as a liquid with a surface temperature very nearly equal to the freezing point of sea water; within this zone Martin and Kauffman (1980) show that the wave decay is linear. Within the liquid region there is also a mean rotary circulation which at the leading edge is of order 0.3 m/sec. Behind the dead zone, the grease ice behaves as a solid and the surface temperature decreases rapidly. Experiments with both plastic chips and oil show that material released on the surface ahead of the dead zone either accumulates at the dead zone or else circulates around as small droplets within the grease ice.

For possible field examples of dead zones, for example, the arrow in Fig. 14-8 marks a detritus line where small ice chunks accumulate and the wave amplitude appears to go to zero; this line is probably the local dead zone and would serve as an accumulation point for oil released on the open water. Again, in Fig. 14-7, we believe that the heads of the plumes marked *c*, *d*, and *e*, where the ice is thicker and there is a wave shadow downwind, also contain local dead zones which would serve as regions of oil accumulation.

To verify these general points, we conducted a specific grease ice oil spill experiment. In this experiment, we grew grease ice to an average depth of about 100 mm at an air temperature of -25°C . We then raised the room temperature to -2°C and generated waves with a period of 0.55 sec, a wavelength of 0.48 m, and an amplitude of 35 mm. These waves were of large amplitude and on the verge of breaking. As soon as the room had warmed up, we poured 200 ml of Prudhoe Bay crude oil onto the water just ahead of the grease ice. We took two kinds of photographs of the spill: first, we placed a camera on a fixed mount above the tank, looking down on the ice at an oblique angle. This camera had a motor drive which took photographs every ten seconds. Second, we used a handheld camera to take photographs of the ice and oil at various positions and times around the tank.

Fig. 14-13 shows the overhead sequential photographs of the spill; the time marked on each photograph is the elapsed time after the start of the spill. The first frame, 0 seconds, shows the pouring of the oil on the surface. The boundary between the open water and the grease ice is visible just below the hand in the photograph, and at the lower left corner of the photograph, the line of bubbles marks the position of the dead zone. These lines are not parallel because the camera is mounted at an oblique angle. The subsequent photographs at 30, 60, and 190 seconds show that the oil was rapidly transported into the

Figure 14-12. Composite figure showing the wave decay and induced circulation within the grease ice. The top shows a side-view photograph of grease ice in our tank; the bottom is a sketch of the induced mean velocity and the final distribution of oil spilled in the ice.

grease ice, then slowly pumped onto the ice beyond the dead zone. Also visible in the photographs at 60 and 190 seconds are the small oil droplets which were produced by the turbulence at the leading edge of the grease ice and are being carried around within the grease ice by the circulation shown in Fig. 14-12. The last two photographs, at 240 and 310 seconds, show that most of the oil is now on the surface beyond the dead zone, with 5-10 percent of the oil circulating as droplets within the grease ice.

For another view of the same experiment, Fig. 14-14 shows several photographs taken with the hand-held camera. Fig. 14-14a shows the oil being poured into the tank, 14-14b shows the appearance of the oil at approximately 10 seconds, and 14-14c the appearance at 250 seconds. In 14-14c, both the wave decay and the pile-up of the oil at the dead zone are clearly visible. Figs. 14-14d and 14-14e, close-ups of the oil droplets, show that these droplets are approximate spheres of 1-3 mm diameter, and circulate within the grease ice without wetting it. Finally, Fig. 14-14f shows the appearance of the oil on the surface after the paddle was turned off at about 30 minutes after the start. In a result similar to Metge's (1978) observations, the photograph shows that because of the grease ice, the oil moves less freely on the surface than it would on open water.

To summarize, the laboratory experiments show that most of the oil ends up on the ice surface beyond the dead zone, with some oil droplets circulating in the grease ice ahead of the dead zone. Our field observations suggest that if oil was spilled in the various polynya zones shown in Fig. 14-5, some would end up on the grease ice surface in the Langmuir convergence zones, some would accumulate in the local dead zones, and a smaller fraction would be emulsified into oil droplets by the breaking waves; these droplets would then circulate around within both the grease ice and the Langmuir rotors. Much of this oil would accumulate downwind of the polynyas in the regions of gray ice as shown south of St. Lawrence Island in Fig. 14-6. Finally, as Pease and McNutt show in their respective chapters (this volume), because these coastal polynyas serve as regions of ice generation for the entire Bering Sea, oil which is released in the polynyas may be carried to the ice edge.

OIL SPILLED IN ICE FLOES OSCILLATING IN A WAVE FIELD

In the field there are two situations in which oil may be associated with ice floes oscillating in a wave field: it may be released under pancake ice, or it may

0 s

30 s

Figure 14-13. Overhead sequential photographs of an oil spill in grease ice.

be released at the ice edge in the 10-20 km zone of floes agitated by ocean swell. In the first situation, Martin et al. (1978) show that pancake ice develops from grease ice, and thus also occurs in the lee-shore polynyas; in the second, Martin and Bauer (this volume) discuss ice-edge floes agitated by ocean swell.

For oil entrainment, it is important in both situations that (1) the floes and cakes exist as separate units with either open water or water and grease ice around them; (2) the incident swell or wind waves cause the floes to oscillate back and forth so that crude oil can be pumped onto their surfaces. For example, in the chapter about the ice edge by Martin and Bauer (this volume), Figs. 12-10a, 12-10b, 12-11a, and 12-11b show ice floes which have been broken by the incident swell, are oscillating in this swell, and have wetted edges caused by the pumping of seawater onto the floe surfaces.

Martin et al. (1978) show that pancake ice forms when grease ice becomes so thick that the circulation within the ice is suppressed; the surface then freezes

into chunks of pancake ice floating over a layer of grease ice. In the field, the pancakes form downwind of the regions of grease-ice formation. On 19 March 1978, we surveyed the ice over a region stretching from Cape Thompson in Kotzebue Sound 90 km south. Within this region, we found an area of at least 100 km² covered with pancakes about 0.5 m in diameter. Fig. 14-15a shows the rough surface appearance caused by rafted pancakes on the site at 67° 19.2'N, 166° 31.3'W, and Fig. 14-15b shows a close-up photograph of the pancakes on the surface. When we cored through the pancakes, we found that they were about 100 mm thick, over a layer of frozen grease ice 285 mm thick, so that the original grease ice thickness was 385 mm. The field evidence suggests that these pancakes grow over large areas under very stormy conditions.

In our previous laboratory study, Martin et al. (1978) found that oil released beneath long linear pancakes came to the ice surface between the oscillating cakes, where the oscillating motion pumped

Figure 14-13, Cont.

60 s

190 s

240 s

310 s

a

b

c

d

Figure 14-14. The oil spill in grease ice. (a) Pouring the oil; (b) the oil after 10 sec; (c) after 250 sec; (d,e) close-ups of the oil droplets; (f) after 30 min.

about 50 percent of the oil onto the ice surface. Because long linear cakes do not appear in nature, we repeated the experiment in a two-dimensional wave-field, so that nearly circular cakes formed in our tank.

We did this experiment in the tank shown in Fig. 14-11. At the side wall of the tank we added a wedge (shown in Fig. 14-16 in plan view) so that the refraction of waves around the wedge generated a two-dimensional wave field. In the experiment, we first grew a layer of grease ice 140 mm thick at an air temperature of -30°C . After the pancakes began to form on the surface, we raised the room temperature to -12°C . The pancakes were 0.1-0.3 m in diameter and 10-20 mm thick. The wave period was 0.91 seconds so that the wavelength was 1.3 m. The wave amplitude at the paddle was 35 mm, and the separation between pancakes oscillated between 0 and 20 mm. Once the pancakes were established, we released 500 ml of Prudhoe Bay crude oil beneath the grease ice through a vertically mounted discharge tube.

Figs. 14-17 and 14-18 show the development of the surface appearance of the oil over 9,160 seconds or about 2.5 hours; the elapsed time from the oil discharge is marked on each photograph in seconds. A 12-inch ruler on the ice just below the spill gives the scale. The photographs show that the oil first appears in the cracks around the pancakes, then the oscillating motion of the cakes drives the oil laterally through the cracks from its point of first appearance and pumps some of the oil onto the floe surfaces.

Figs. 14-19a and b show a lateral view of the spill at about 60 sec and at 1,660 sec respectively. These lateral views show that the ridge heights around the cakes vary from 0 to 30 mm, and that the oil is pumped over the rims onto the interior of the pancake.

We also found that, although the grease ice in the cracks around the pancakes inhibited the oil movement through the crack system, the oil was pumped laterally through the cracks with what appeared to be the higher fractions, or lighter-colored oil, running ahead. As the final photograph in Fig. 14-16 shows, at the end of the experiment, the oil was contained with considerable density variation within a region measuring $1 \times 0.6 \text{ m}^2$. This means that on the average, 1 ml of oil covered 12 cm^2 .

On the day after the experiment, we scraped up the oil from the ice surface and from the cracks where the oil was accessible, to recover the following amounts of oil: from the surface, 32 ml; from the cracks, 90 ml. This meant that a total of 120 ml of

Figure 14-15. Pancake ice in the field. (a) Surface appearance of large pancake-covered area; (b) close-up of the pancakes with chisel for scale.

Figure 14-16. Plan-view sketch of tank used for pancake ice experiments.

oil was accessible to cleanup from the surface, or about 25 percent of the total spilled.

To search for the rest of the oil, we then cut up the ice into vertical sections around the oil spill. In a schematic drawing, Fig. 14-20 shows that the oil beneath the surface occurred on the pancake bottoms, deep within the cracks, and, in the form of small droplets, within the grease ice. The importance of the grease ice, then, is that it retards both the rise of oil to the surface and its lateral spreading within the cracks. In summary, in the case of an oil spill beneath a wind-agitated field of pancakes, we can expect that a sizeable fraction of the oil will be pumped onto the ice surface, and that the remainder will be bound up in the grease ice/pancake ice system. Second, the observations of swell propagation through the fields of ice floes near the ice edge and the observations of wetted surfaces on these floes strongly suggest that a similar pumping of oil onto the floe surfaces will occur in this region.

OTHER TRANSPORT PROCESSES

Finally, we consider the possible large-scale movement of oil by the general ice circulation in the Bering Sea described in the chapters by McNutt, Pease, and Martin and Bauer (this volume). Fig. 14-21 is a schematic drawing of the large-scale winter ice motion during periods of northeast winds. Proceeding from the upwind direction the figure shows the production of grease ice in the lee-shore polynyas and the general southward advection of the large ice floes. As these large floes near the ice edge, the propagation of ocean swell into the pack breaks them into smaller floes (Martin and Bauer, this volume). As these continue to approach the edge, the action of wind and swell causes rafting and ridging, so that sails as high as 1 m and keels as deep as 5 m form.

This increase in aerodynamic roughness means that the wind moves these floes to the southwest faster than those of the unroughened pack. As Bauer and Martin show (this volume), at the ice edge these floes are organized into the bands observed on satellite photographs like Fig. 12-5; these bands move southwest at velocities of about 4 percent of the wind speed at 25° to the right of the wind. As they move into warmer water, the bands disintegrate, thereby exposing new pack ice to be broken by the swell.

To summarize this process, the wind stress causes the interior pack ice to move southwest to the edge from the lee-shore polynyas. Pease shows that this movement from the polynyas to the ice edge takes 30-60 days. As the ice approaches the edge, the wave action roughens the ice so that it forms into bands which move southwest at a higher velocity than the rest of the pack. This southwest motion continues until the ice melts. This "conveyor belt" then, may transport oil which is spilled in the northern Bering Sea to a location well south of the ice edge. Also, if oil were spilled at the ice edge, this ice motion could capture the slick and spread it into open water farther south.

The laboratory and field studies on oil pollution and sea ice suggest that much of the spilled oil will accumulate on the water and ice surface. Because our laboratory experiments are of short duration, however, we do not yet have a good idea of how the oil will evolve once it is on the surface. Field studies of oil spills in temperate water—for example, the 1976 *Argo Merchant* (Grose and Matson 1977) and the 1979 IXTOC-1 blowout—show that, because of the combination of wind and wave mixing with evaporation and solar photo-oxidation, the oil forms a mousse-like emulsion with seawater, which then forms into pancakes with horizontal dimensions of 1-100 m and thicknesses of 0.05-0.3 m. If similar emulsions form in polar ocean spills, they may not interact with the ice in the same way as the crude oil and diesel spills described in this chapter.

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20 s

60 s

0 s

40 s

100 s

320 s

Figure 14-17. Overhead sequential photographs of the oil and pancake experiment. The elapsed time from the oil release is shown in seconds on each photograph.

2020 S

5740 S

1270 S

2740 S

9160 s

7540 s

Figure 14-18. The continuation of the sequence in Fig. 14-17.

on running the oil-spill experiments; we greatly appreciate his help. This study was supported by the Bureau of Land Management through interagency agreement with the National Oceanic and Atmospheric Administration, under which a multiyear program responding to needs of petroleum development of the Alaskan continental shelf is managed by the Outer Continental Shelf Environmental Assessment Program (OCSEAP) Office. This is publication number 1120 of the Department of Oceanography, University of Washington, Seattle.

Figure 14-19. Side-view photographs of the oil spill shown in Figs. 14-17 and 14-18. (a) 60 sec.; (b) 1660 sec.

Figure 14-20. Schematic side-view drawing of the oil location within a pancake ice/grease ice mixture.

Figure 14-21. A sketch showing the general ice motion and deformation during periods of northeast winds.

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Section III

Geology and Geophysics

C. Hans Nelson, editor

Sedimentary Processes and Potential Geologic Hazards on the Sea Floor of the Northern Bering Sea

15

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ABSTRACT

A dynamic environment of strong bottom currents, storm waves, and gas-charged sediment on the shallow sea floor of the northern Bering Sea creates several potential geologic hazards for resource exploration. Thermogenic gas seeps, sea-floor gas cratering, sediment liquefaction, ice gouging, scour-depression formation, coastal and offshore storm surge and associated deposition of storm-sand, and movement of large-scale bedforms all are active sedimentary processes in this epicontinental shelf region.

Interaction between the processes of liquefaction and the formation of shallow gas pockets and craters, scour depressions, storm-sand deposits, and slumps results in sediment instability. Liquefaction of the upper 1-3 m of sediment may be caused by cyclic storm-wave loading of the Holocene coarse-grained silt and very fine grained sand covering Norton Sound. The widespread occurrence of gas-charged sediment with small surficial craters (3-8 m in diameter and less than 1 m deep) in central Norton Sound indicates that the sea-floor sediment is periodically disrupted by escape of biogenic gas from the underlying peaty mud. During major storms, liquefaction may not only help trigger crater formation but also magnify erosional and depositional processes that create large-scale scour areas and prograde storm-sand sheets in the Yukon prodelta area.

Erosional and depositional processes are most intense in the shallower parts of the northern Bering Sea and along the coastline during storm surge flooding. Ice gouges are numerous and ubiquitous in the area of the Yukon prodelta, where the sediment is gouged to depths of 1 m. Though much less common than in the prodelta, ice gouges are present throughout the rest of the northern Bering Sea where water depths are less than 20 m and at times where water is as much as 30 m deep. In the Yukon prodelta area and in central Norton Sound, where currents are constricted by shoal areas and flow is made turbulent by local topographic irregularities (such as ice gouges), storm-induced currents have scoured large (10-150 m in diameter), shallow (less than 1 m deep) depressions. The many storm-sand layers in Yukon prodelta mud show that storm surge and waves have generated bottom-transport currents that deposit layers of sand as thick as 20 cm as far as 100 km from land. Storm-surge runoff may reinforce the strong geostrophic currents near Bering Strait, causing intermittent movement of even the largest sand waves (10-200 m wavelength, to 2 m height).

INTRODUCTION

Studies of potential geologic hazards on the Norton Basin sea floor in the northern Bering Sea have been conducted by the U.S. Geological Survey (U.S.G.S.) in the course of evaluating oil and gas lease tracts preparatory to Outer Continental Shelf (OCS) leasing. The data base for this evaluation included 9,000 km of high-resolution geophysical tracklines (Nelson et al. 1978b, Thor and Nelson 1978, Larsen et al. 1979a), 1,000 grab samples, 400 box cores, and 60 vibracores; in addition, hundreds of camera, hydrographic, and current-meter stations have been occupied during the past decade by U.S.G.S., National Oceanic and Atmospheric Administration (NOAA), and University of Washington oceanographic vessels (Figs. 15-1 and 15-2).

The northern Bering Sea is a broad, shallow epicontinental shelf region covering approximately 200,000 km² of subarctic sea floor between northern Alaska and the U.S.S.R. The shelf can be divided into four general morphologic areas: (1) the western part, an area of undulating, hummocky relief formed by glacial gravel and transgressive-marine sand substrate (Nelson and Hopkins 1972); (2) the southeastern part, a relatively flat featureless plain with fine-grained transgressive-marine sand substrate (McManus et al. 1977); (3) the northeastern part, a complex system of sand ridges and shoals with fine-to medium-grained transgressive sand substrate (Nelson et al. 1978a); and (4) the eastern part, a broad, flat marine reentrant (Norton Sound) covered by Holocene silt and very fine sand (Nelson and Creager 1977). A detailed discussion of bathymetry

Figure 15-1. Sampling stations for U.S. Geological Survey research in northern Bering Sea between 1967 and 1978.

and geomorphology of the northern Bering Sea is given by Hopkins et al. (1976) (Fig. 15-3).

The northern Bering Sea is affected by a number of dynamic conditions: winter sea ice, sea-level setup, storm waves, and strong currents (geostrophic, tidal, and storm). The sea is covered by pack ice for about half the year, from November through May. A narrow zone of shorefast ice (sea ice attached to the shore) develops around the margin of the sea during winter months. Around the front of the Yukon River Delta, shorefast ice extends to 40 km offshore (Thor et al. 1978). During the open-water season, the sea is subject to occasional strong northerly winds, and in the fall strong south-southwesterly winds cause high waves and storm surges along the entire west Alaskan

coast (Fathauer 1975). Throughout the year, there is a continual northward flow of water with currents intensifying on the east side of strait areas (Coachman et al. 1975). Although diurnal tidal ranges are small (less than 0.5 m), strong tidal currents are found in shoreline areas and within central Norton Sound (Fleming and Heggarty 1966, Cacchione and Drake 1979a).

This chapter reviews basic sedimentary processes of this epicontinental shelf region and discusses certain potential geologic hazards related to these processes: thermogenic gas seepage, biogenic gas saturation of sediment and cratering, sediment liquefaction, ice gouging, current scouring, storm-sand deposition, and mobile bedform movement (Fig.

Figure 15-1a. Cross-hatched area of Fig. 15-1, showing closely spaced sampling grid offshore at Nome.

15-4). These geologic hazards may pose problems for the future development of offshore resources in Norton Basin.

SEDIMENTARY PROCESSES

Yukon Delta processes

The Yukon River drains an area a little less than 900,000 km², providing a water discharge of approximately 6,000 m³/sec and a sediment load of 70-90 million mt per year (Dupré and Thompson 1979, Cacchione and Drake 1979a). The sediment load, almost 90 percent of all sediment entering the Bering Sea, is composed mainly of very fine sand and coarse silt with very little clay.

The Yukon Delta plain, like many deltas, is fringed by prograding tidal flats and distributary mouth bars. The delta front and prodelta are offset from the prograding shoreline by a broad platform (referred to as a subice platform) 30 km at its widest reach. This platform appears to be related to the presence of shorefast ice that fringes the delta for half the year. The term "delta front" describes the relatively steep margin of the offshore delta environment characterized by rapid deposition of sediment in water 2-10

m deep. The prodelta, an area of extremely gentle slopes, marks the distal edge of the deltaic sediments extending as far as 100 km offshore.

Processes on the Yukon Delta and offshore operate under seasonal regimens (Dupré and Thompson 1979). The ice-dominated regimen begins with freeze-up in late October or November. Shorefast ice extends 10-40 km offshore, where it terminates in a series of pressure ridges and shear zones formed by the interaction of shorefast ice with the highly mobile seasonal pack ice.

River breakup, typically in May, marks the beginning of the river-dominated regimen. Once the shorefast ice melts or drifts offshore, sedimentation is dominated by normal deltaic processes under the influence of the high discharge of the Yukon River.

Increasingly frequent southwest winds and waves associated with major storms during late summer mark the beginning of the storm-dominated regimen. High wave energy and decreasing sediment discharge from the Yukon cause considerable coastal erosion and reworking of deltaic deposits.

Coastal storm surge

In November 1974 a severe storm moved from

Figure 15-2. Geophysical trackline surveys for U.S. Geological Survey research in northern Bering Sea 1967-78.

Figure 15-3. Generalized bathymetry of northern Bering Sea in 10-m contour intervals.

southwest to northeast across the Bering Sea. Peak winds were 111 km per hour from the south, and nearshore waves were reported to be 3-4 m in height (Fathauer 1975). Coastal flooding extended from Kotzebue Sound (north of Bering Strait) to just north of the Aleutian Islands (Fathauer 1975). The maximum sea-level setup, measured by the elevation of debris lines along the coast of Norton Sound, ranged from 3 to 5 m above mean sea level (Sallenger et al. 1978). During this storm, extensive inland flooding occurred and erosion of coastal bluffs 2-5 m high took place near Nome. Irregular landward erosion, as much as 18 m, occurred west of Nome, where bluffs are 3-5 m high. East of Nome, where bluffs are 1.5-2 m high, landward erosion was as much as 45 m. The

water level in the Norton Sound area reached its peak on 12 November, when as much as 2 m of water was standing in the village of Unalakleet and the static high-water line at Nome was 4 m above mean low water.

Storm currents

The transport of sediment in Norton Sound and Norton Basin can be described in terms of distinct quiescent and storm regimes (Cacchione and Drake 1979a). The quiescent regime is characterized by generally low levels of sediment transport caused mainly by tidal currents. Fine silt and clay move as "wash load," and bedload transport is negligible except in shallow areas where surface waves become

Figure 15-4. Potentially hazardous areas of northern Bering Sea (from Thor and Nelson 1979).

dominant. Current speeds in this regime are no greater than 30 cm/sec.

Calm weather conditions prevail for about 90 percent of the year in the northern Bering Sea, and less than 50 percent of the sediment transport takes place under these conditions (Cacchione and Drake 1979a). Norton Sound is commonly exposed to strong southerly and southwesterly winds generated by low-pressure weather systems in September, October, and November. A two-day storm in September 1977 transported sediment equal to the transport that would occur during four months of quies-

cent conditions. Current speeds were as much as 70 cm/sec during this storm.

Graded storm-sand layers to 20 cm thick (Nelson 1977) occur in sea-floor strata of the northern Bering Sea, widespread evidence of major storm-surge events. The effects of storm surge are intensified by two factors: extremely shallow water depth (less than 20 m), and strong bottom return currents that may move large amounts of sediment northward to the Chukchi Sea. The thickness of Holocene sediment in Norton Sound relative to Holocene sediment input from the Yukon River indicates that significant amounts of

sediment have been resuspended and transported out of Norton Sound (Nelson and Creager 1977). About 10 percent of the Yukon River input into Norton Sound may be carried as suspended sediment through the Bering Strait into the Chukchi Sea under non-storm conditions (Cacchione and Drake 1979a). As much as 40 percent of Holocene sediment discharged from the Yukon River appears to be missing from Norton Sound; this difference of 30 percent may have been resuspended and transported during storms (Cacchione and Drake 1979a).

Storm currents not only resuspend and transport massive amounts of suspended sediment, but also appear to move large amounts of sand in bedload transport for considerable distances offshore. Graded storm-sand layers are extensive throughout southern Norton Sound; their thickening toward the Yukon subdelta, the apparent source region, suggests massive movement of bedload sediment away from the delta toward the adjacent offshore region during storms (Nelson 1977).

Wave effects

Waves and wave-induced currents are the dominant sedimentological agents on the inner shelf of northwestern Norton Sound and in the approaches to Bering Strait (Hunter and Thor 1979). Sedimentary features common to both areas include sand and gravel patches and ribbons, wave ripples, sand waves, and ice gouges (Hunter and Thor 1979).

Wave ripples with spacings to 2 m are common in both the Port Clarence and Nome areas in zones where sediment is well sorted and grain size ranges from coarse sand to pebbly gravel. Ripples in the Port Clarence area trend northwest-southeast and can be explained as the result of storm waves from the southwest Bering Sea. Trends of ripples in the Nome area indicate dominant wave activity from south to southwest.

Ribbons of sand and gravel are well developed near the entrance to Port Clarence. These bedforms may be produced by wave action or by wave-induced net water motion in the direction of wave propagation.

A rich assemblage of depositional and erosional features, both wave-formed and current-formed, occupies the floor in shallow water close to the southern shore of Seward Peninsula. Wave-formed features are more common; some of the current-formed features imply considerable sediment transport by strong bottom tidal currents.

Only the broad patterns of wave and current movement in southwestern Norton Sound are known. The major wave trains originate in the southern Bering Sea; waves move northward and refract

clockwise around protruding Yukon shoals. Smaller waves with shorter periods are generated by northeasterly winds and move southwestward.

Liquefaction

The Yukon River sediment that covers most of the bottom of Norton Sound (McManus et al. 1977) is primarily silt with considerable amounts of very fine sand in some areas and a generally minor content of clay-size material. The sediment thickness is usually less than 3 m, except near the Yukon Delta, where accumulations are as thick as 10 m (Olsen et al. 1979). The material is generally dense; there are zones of relatively loose sediment (material of low density) in gas-charged areas. In the delta areas sampled by 6-m vibracores, relatively loose zones of sediment were observed above and between dense layers.

Freshwater peaty mud beneath Yukon marine silt is somewhat over-consolidated and contains substantial amounts of organic carbon and gas. The presence of gas indicates that the pore pressure in the peaty muds may be high. If it is, the strength of the material could be low despite its highly consolidated state.

The dominantly coarse-silt to fine-sand texture of the material, the occurrence of loose sediment zones, and theoretical calculations utilizing GEOPROBE cyclic wave loading data (Olsen et al. 1979, Clukey et al. 1980) indicate that Yukon prodelta sediment in southwestern Norton Sound is susceptible to liquefaction. Liquefaction of the prodelta deposits may be caused by cyclic loading resulting from exposure of the Yukon prodelta to large storm waves from the southwest. Water depths are shallow enough for much of the wave-generated surface energy to be imparted to the bottom sediment, so that the upper 1-2 m of sediment may be liquefied during extreme storm-surge events (Clukey et al. 1980). This liquefaction of prodelta sediment may influence storm-sand transport, formation of sediment depressions, and gas cratering.

Ice scour

Ice on the Bering shelf scours and gouges surficial sediment of the sea floor (Fig. 15-5). The annual ice cover in this subarctic setting is generally thin (less than 2 m); thick ice capable of gouging forms where pack ice collides with and piles up against stationary shorefast ice, developing numerous pressure ridges (Thor and Nelson, this volume). A wide, well-developed shear zone forms in southwestern Norton Sound as ice moving southward from the northeast Bering Sea and westward along southern Norton Sound converges in the shallow water of

Figure 15-5. Distribution and density of ice gouging, direction of movement of pack ice, and limits of shorefast ice in northern Bering Sea (from Thor and Nelson 1979).

the Yukon prodelta. Consequently, numerous zones of pressure ridges are formed. In this region, where the water is 10-20 m deep, the density of ice gouges is highest. Gouges are found in water to 30 m deep, and furrows are as much as 1 m deep. Ice gouging affects the sea floor under shorefast areas only minimally or not at all (Thor and Nelson, this volume).

Current scour depressions

Zones of large flat-floored depressions in Norton Sound occur mainly in two areas: west of the Yukon prodelta and 50 km southeast of Nome on the flank of a broad shallow trough (Fig. 15-6) (Larsen et al. 1979a). These features range from individual more or less elliptical depressions 10-30 m in diameter to large areas with irregular margins, 80-150 m in diameter.

The depressions are 60-80 cm deep (Larsen et al. 1979b).

Bottom current speeds in depression areas are 20-30 cm/sec under nonstorm conditions and were measured at 70 cm/sec during a typical autumn storm (Cacchione and Drake 1979a). Both zones of depressions are on flanks of gently sloping shoals, where strong tidal or geostrophic currents shear against the slopes. Small-scale ripple bedforms are associated with depression areas and mean grain size ranges from 4 to 4.5 ϕ (0.063-0.044 mm). Depressions in the Yukon Delta area are associated with extensive ice gouging. The gouge furrows commonly expand into large shallow depressions (Larsen et al. 1979 and Thor and Nelson, this volume).

Experiments in flumes containing fine sand and silt have shown that currents flowing over an obstruction will scour material immediately downcurrent

Figure 15-6. Location of scour depressions, extensive scour and ripple zones, and strong bottom currents in Norton Sound, showing area of storm sand deposition (modified from Larsen et al. 1979b).

from the obstruction (Young and Southard 1978). The large scour depressions observed in Norton Sound may be a characteristic erosional bedform developed during storms when strong currents and high wave energy are focused on silt-covered slopes where local topographic disruptions set off flow separation and downcurrent scour.

Sandwave dynamics

Strong geostrophic currents prevail throughout much of the northern Bering Sea, particularly where westward land projects into the northward flow, as in the eastern Bering Strait area (Fleming and Heggarty 1966, Coachman et al. (1975). In such regions large bedforms develop and migrate, forming an unstable sea floor (Nelson et al. 1978a). These large bedforms include large-scale sand waves 1-2 m

high with wavelengths to 200 m, and small-scale sand waves 0.5-1.0 m high with wavelengths of 10 m. They occupy the crests and some flanks of a series of linear sand ridges 2-5 km wide and as much as 20 km long between Port Clarence and King Island.

Sand-wave movement and bedload transport take place during calm weather (Nelson et al. 1978a), but maximum change apparently occurs when severe southwesterly storms generate sea-level setup in the eastern Bering Sea that enhances northerly currents. In contrast, strong north winds from the Arctic reduce the strength of the northerly currents and thereby arrest bedform migration.

Sediment gas charging

The distribution of acoustic anomalies suggests that almost 7,000 km² of sea floor in Norton Sound

and Chirikov Basin is underlain by sediment containing gas (biogenic and/or thermogenic) sufficient to affect sound transmission through these zones (Holmes 1979a). Core-penetration rates (Nelson et al. 1978c, Kvenvolden et al. 1979c) and sediment samples from 2-6 m vibracores confirm gas saturation of near-surface sediment at several locations characterized by acoustic anomalies. The isotopic compositions of methane at four of the sites range from -69 to -80 ‰ ($\delta^{13}\text{C}_{\text{PDB}}$) (Kvenvolden et al. 1978, 1979b). This range of values clearly indicates that the methane is formed by microbial processes, possibly operating on near-surface Pleistocene peat deposits that underlie Holocene deposits throughout the northern Bering Sea.

At one site in Norton Sound, near-surface sediment is apparently charged with CO_2 actively seeping from the sea floor accompanied by less than one percent hydrocarbon gases (Kvenvolden et al. 1979c). Methane in this gas mixture has an isotopic composition of -36 ‰, a value suggesting that it is derived mainly from thermal processes, probably operating at depth in Norton Basin. Geophysical evidence indicates that the hydrocarbon gases migrate into the near-surface sediments along a fault zone (Nelson et al. 1978c). Subbottom reflector terminations on continuous seismic profiles near the fault zone outline a large zone of anomalous acoustic responses about 9 km in diameter at a depth of 100 m, caused by a thick subsurface accumulation of gas. Gas geochemistry and extensive voids due to gas expansion in vibracores suggest a high degree of gas saturation at the seep site (Kvenvolden et al. 1979b).

Biogenic gas cratering

Small circular pits on the sea floor are found over a 20,000-km² area of central and eastern Norton Sound (Fig. 15-7). The craters in the northern Bering Sea are young, as shown by their presence within modern ice-gouge grooves and by the fact that relict buried craters have not been observed in seismic profiles (Nelson et al. 1979b). These craters range from 1 to 10 m in diameter, averaging 2 m, and are probably less than 0.5 m deep. They are associated with many acoustic anomalies observed on seismic profiles and with subsurface Pleistocene peaty mud that is commonly saturated with biogenic methane (Holmes 1979b, Kvenvolden et al. 1979a, Nelson et al. 1979a). The extensive reflector-termination anomalies and peat with a high gas content in east-central Norton Sound suggest that gas-charged sediment may be the cause of crater formation.

Two basic mechanisms for gas venting are possible. First, continuous local degassing may maintain craters

as active gas vents on the sea floor. Second and more likely, gas may be intermittently vented, particularly during severe storms when near-surface sediment may liquefy.

The occurrence of surface craters in overlying marine sediment and the presence of high quantities of methane trapped beneath cohesive marine mud in Norton Sound suggest that gas venting may be episodic in this lithologic setting. Absence of craters in the noncohesive near-surface fine-to-medium sand and gravel of Chirikov Basin indicates that gas probably diffuses gradually through this more porous sediment that overlies the peaty mud. Further evidence for intermittent venting of gas is the broad, shallow shape of the craters, unlike the deep, conical, actively bubbling vents of the thermogenic seep. Lack of methane in bottom water also suggests that the craters are not continually active vents.

POTENTIAL GEOLOGIC HAZARDS

Thermogenic gas cap

The extent of active gas seepage into the water column and gas saturation in near-surface sediment above a thick sediment section with acoustic anomalies suggest a possible hazard for future drilling activity in the thermogenic gas seep area south of Nome. Artificial structures penetrating the large gas accumulation at 100 m or intersecting associated faults that cut the gas-charged sediment may provide direct avenues for uncontrolled gas migration to the sea floor.

Shallow gas pockets

Gas-charged sediment creates potentially unstable surficial-sediment conditions in Norton Sound. Approximately 7,000 km² of Norton Sound is underlain by acoustic anomalies with potential shallow gas pockets everywhere except under the Yukon prodelta (Holmes 1979a, Nelson et al. 1979a). Pipelines built across areas of these potential gas pockets may be damaged by stress induced by the unequal bearing strength of gas-charged and normal sediment, particularly if the near-surface sediment is undergoing liquefaction caused by cyclic loading of storm waves. The gas saturation and lateral and subsurface extent of any shallow gas pockets will have to be detailed in any site investigations for platforms or pipelines.

Gas craters

Gas craters cover a large area of north-central Norton Sound. During nonstorm conditions, near-surface gas in this area may be trapped by a layer 1-2

Figure 15-7. Distribution and density of biogenic gas-generated craters on sea floor of Norton Sound, showing isopachs of Holocene mud derived from the Yukon River and deposited since Holocene postglacial sea-level rise (from Thor and Nelson 1979).

m thick of impermeable Holocene mud. We postulate that the gas escapes during periodic storms, forming craters at the surface. The storm processes initiate rapid changes in pore-water pressures because of sea-level setup, seiches, erosional unloading of covering mud, and possible sediment liquefaction from cyclic wave loading (Clukey et al. 1980). Gas venting and sediment craters or depressions which seem to form during peak storm periods may be a potential hazard to offshore facilities because of the rapid lateral changes in bearing strengths and collapse of sediment which form the craters. Sediment collapse may also expose pipelines to ice-gouging hazards. During nonstorm conditions, the upper several meters of sediment at many locations has reduced shear strength because of the near-surface gas saturation and presence of peat layers. Choosing locations for structures will require extensive testing of the substrate to determine the extent and activity of gas cratering at a given site.

Liquefaction

The assessment and prediction of sea-floor stability are affected by the possibility that a sedimentary deposit will liquefy under cyclic loading and behave as a viscous fluid. The liquefaction potential of Norton Sound sediment is great in central Norton Sound and in the vicinity of the western Yukon prodelta (Clukey et al. 1980, Olsen et al. 1979). Possible causes of liquefaction include upward migration of gas from thermogenic and biogenic sources, earthquakes, and ocean waves. Bottom features that may be caused in part by liquefaction include scour depressions and much sediment cratering where Yukon sediment is thin.

Loss of substrate support by sediment liquefaction must be considered in the construction of pipelines, drilling platforms, and other types of structures resting on the sea floor. Full assessment of this problem requires extensive studies of *in-situ* pore

pressure, gas saturation, and wave cyclic loading during storms.

Ice scour

The maximum intensity of ice gouging occurs in central Norton Sound at depths of 10-15 m in an area surrounding the Yukon Delta. The remaining area of Norton Sound, where depths are less than 10 m or more than 20 m, has a low density of gouging, or none at all. Special studies of nearshore areas off Nome and Port Clarence were made when they became potential centers for commercial development and activity. Offshore of Nome, the focal point for logistics in the northern Bering Sea because it is an area of ice divergence, only a few gouges were found in water more than 8 m deep (Thor and Nelson, this volume). Several gouges were found at the northern end of Port Clarence spit and inside the tidal inlet, but again none occurred in water less than 8 m deep.

Ice gouging presents some design problems and potential hazards to installations in or on the sea floor. Pipelines and cables should be buried at a depth that allows for maximum ice gouging of 1 m, plus a safety factor for combined effects with current scour around the western Yukon prodelta front or gas cratering in central Norton Sound.

Current scour depressions

The highest density of scour depressions in Norton Sound is in two areas: west and northwest of the Yukon delta, and southeast of Nome (Fig. 15-6). In areas of high density, artificial structures that disrupt current flow may cause extensive erosion of Yukon-derived silt or very fine grained sand and create potentially hazardous undercutting of the structures. Even buried structures such as pipelines may be subject to scour because strong currents can greatly broaden and deepen naturally occurring ice gouges, thus exposing the structures. The abundance and lateral extent of scour depressions are greatest where they occur with ice gouging in the Yukon Delta areas. Replicate surveys have shown that scour depressions recur annually. Full assessment of this geologic hazard requires long-term current monitoring in specific localities of scour to predict current intensity and periodicity, especially during severe storms: measured current speeds have increased more than 100 percent even under moderate storm conditions (Cacchione and Drake 1979b).

Mobile bedforms

Large migrating bedforms form an unstable sea floor in the area west of Port Clarence. The actual rates of bedform movement are not known, but development and decay of sand waves up to 2 m in height has been observed during a one-year period. Pipelines could be subject to damaging stress if free spans developed where the structure crossed such areas of migrating sand waves 2 m high.

Studies to this time indicate that potential for the most extreme scour exists in regions of sand ribbons and gravel plus shell pavement within the strait. Sea floor relief changes most rapidly in the Port Clarence sand-wave area, where the scour in sand-wave troughs may reach depths of 2 m. Replicate surveys have shown that such scour may occur each year in some areas of the Port Clarence sand-wave field. Long-term monitoring of currents and bedform movement is particularly important in determining actual rates of change in this area, the only large natural harbor on the Alaskan coast north of the Aleutians.

Coastal and offshore storm-surge hazards

The northern Bering Sea has a history of major storm surges accompanied by widespread changes in sea-floor sedimentation (Cacchione and Drake 1979a, Fathauer 1975, Nelson 1977); these changes complicate maintenance of sea-floor installations and mass transport of pollutants. The November 1974 storm was the most intense measured in historic time, but storms of 1913 and 1946 also caused considerable damage (Fathauer 1975). Severe storms, such as that of November 1974, have caused extensive flooding along the Norton Sound coast between Nome and Unalakleet and on the St. Lawrence Island coast (Sallenger et al. 1978). At Nome, storm surge and waves overtopped a sea wall, causing damage reported at nearly 15 million dollars (Fathauer 1975). The periodicity and intensity of storm surges will have to be carefully studied in planning whether and where pipelines should come ashore in this area.

Rapid sedimentation of thick storm-sand layers (15-20 cm) is a problem in the Yukon Delta area. Pipelines, offshore facilities, and other structures impeding the erosion, transport, and redeposition of sediment in southern Norton Sound will require careful design. Accurate monitoring of the storm-surge process will require long-term deployment of an array of current meters and tide gauges in the northern Bering Sea.

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The Ice-dominated Regimen of Norton Sound and Adjacent Areas of the Bering Sea

16

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ABSTRACT

The patterns of ice formation, movement, and deformation in Norton Sound and adjacent areas of the Bering Sea were studied using LANDSAT and NOAA satellite imagery for the years 1973-77. The results demonstrate not only the marked seasonality of marine processes, but also the significant role of bathymetric and meteorologic conditions in controlling ice movement in this high-latitude, epicontinental sea.

The ice-dominated regimen of the Norton Sound region begins with ice freezeup in October and lasts through the winter to spring breakup in May. Freezeup begins as temperatures drop in early fall and ice starts to accumulate around the Yukon Delta; during that time water and sediment discharge from the Yukon River becomes insignificant. Oceanographic currents are relatively ineffective in transporting ice during the winter months, as strong northerly winds generally control ice movement in the winter phase. Consequently, ice divergence from the northern coast of Norton Sound and ice convergence along the northern margin of the Yukon Delta front are common throughout the winter phase. Ice ridging and associated gouging result from the compaction of shorefast ice along the northern delta front. Shearing and gouging also occur along the western margin of the delta where pack ice, pushed southward by the winds, shears past the shorefast ice boundary. Periods of rapid advection of pack ice from the Chukchi Sea through the Bering Strait are also common in the winter phase. Such events cause ice floes to move as much as 45 km per day, usually along a relatively narrow band on the eastern side of the Bering Sea.

During May, winds become predominantly offshore and warming temperatures begin to melt the ice and snow in the interior. The higher temperatures and increased discharge from the rivers trigger ice breakup along the coast. Northward-flowing water currents aided by offshore winds carry ice away from the delta. By June the high sediment influx from the Yukon River dominates the coastal processes in the delta region.

INTRODUCTION

The prospect of oil and gas exploration in Norton Sound (Fig. 16-1) has focused increased attention on the marine processes that characterize the region.

Figure 16-1. Location of study area.

Because these processes vary systematically throughout the year, it is possible to recognize ice-dominated, river-dominated, and storm-dominated regimens (Fig. 16-2); each regimen consists of a characteristic suite of geologic processes and associated potential geologic hazards (Dupré and Thompson 1979).

The purpose of this paper is to discuss the ice-dominated regimen of Norton Sound from freezeup,

Figure 16-2. Diagram illustrating the seasonality of processes in the Yukon Delta/Norton Sound region of the Bering Sea. Sources of data include: (a) Summary of average monthly temperatures at Unalakleet, 1941-70 (NOAA); (b) Summary of ice observations for Yukon Delta region from Brower et al. (1977); (c) Frequency of major low-pressure centers in the northern Bering Sea region, from Brower et al. (1977); and (d) Discharge of the Yukon River at Kaltag, 1962 (U.S.G.S. Water Resources Data).

typically in late October or early November, to breakup in May (Fig. 16-3). Of particular interest is the extent and variability of shorefast ice, the patterns and rates of movement of seasonal pack ice, and the weather conditions under which such movement occurs. Most of this work has been based on the interpretation of satellite imagery, but we hope that this study can provide background for more detailed studies based on ground monitoring and *in-situ* measurements.

Figure 16-3. Three phases of the ice-dominated regimen.

Previous work

Most of the work on ice dynamics and the possible effect of ice on offshore petroleum development has been concentrated in Arctic regions such as the Beaufort Sea; relatively little work has been published to date on ice movement and morphology in sub-

arctic regions such as the Bering Sea and Norton Sound. Muench and Ahlnäs (1976) were the first to use satellite imagery to study regional patterns of ice movement in the northern Bering Sea; the relatively low resolution of their imagery (NOAA weather satellite data) and the relatively short period of record (March to June 1974), however, limited the utility of their work. Shapiro and Burns (1975) used higher-resolution LANDSAT imagery to document a short-lived ice deformation event just to the north of the Bering Strait. Stringer (1977, 1978) mapped a variety of ice-related features in the Beaufort, Chukchi, and Bering Seas with the use of LANDSAT imagery. Similarly, Dupré (1978) used LANDSAT imagery to study the complex interrelationships between ice and patterns of deltaic sedimentation associated with the Yukon Delta. This present study is designed to expand on these previous studies, emphasizing the patterns and rates of ice movement in Norton Sound. In doing so, the study also provides information to aid in the explanation and extrapolation of patterns of ice gouging in Norton Sound as described by Thor et al. (1978).

Methods of study

The data were compiled from imagery acquired by the Multispectral Scanner system of the LANDSAT and NOAA-2, 3, and 4 satellites. Meteorologic data were also taken from daily surface synoptic charts from the National Climatic Center in North Carolina. The extent of shorefast ice and ice floes was mapped from LANDSAT images (1:1,000,000) on acetate overlays that were superimposed on standard bathymetric base maps of the northern Bering Sea. Overlays for successive days could be superimposed to chart the movement of particular ice floes over an approximate 24-hour period. Images were registered with respect to landforms in order to map the positions of the ice floes on successive days. In general, sea ice movement cannot be accurately monitored by referencing the scenes to coordinates, because coordinates provided on the margin of LANDSAT images allow for only approximate registration (Colvocoresses and McEwen 1973). According to Colvocoresses and McEwen (1973), the systematic, root mean square error of position for points on the satellite images ranges from 200 to 450 meters, with no detectable additional error associated with image duplication. Since the sea ice is moving on the order of kilometers per day, this error should be considered insignificant for the purposes of this paper.

The LANDSAT images were band 5 and 7, 9 × 9-inch positive prints. The images and the base map seemed to be perfect overlays. The band-5 images

proved to be most useful in defining nilas ice (a thin, elastic crust), whereas band 7 was more useful for defining the sea-ice boundary and delineating pack-ice floes and areas of shorefast ice. There is no distinction made between newly formed ice and open water on most maps in this paper, because of the difficulty in distinguishing the two.

The NOAA 10 × 10-inch satellite photoprints (infrared) furnished only general meteorological information. They did provide a good overview of ice movement, but were not used for detailed measurements. Wind patterns and weather systems could be observed from NOAA imagery, but wind velocity and directional information was obtained from daily surface synoptic charts. Although some weather station readings are influenced by local orographic conditions, most of the information is believed to be useful for the purposes of this study.

ICE-DOMINATED REGIMEN

The pattern of ice formation, movement, and deformation in the Norton Sound region is significantly affected by the nearshore morphology of the Yukon Delta. The Yukon River has formed an ice-dominated delta characterized by a broad, shallow sub-ice platform crossed by sub-ice channels that extend as far as 25 km beyond the major distributaries (Fig. 16-4). The platform is characterized by relatively stable shorefast ice for much of the year, whereas the sub-ice channels have more dynamic ice and sediment movement, particularly during breakup. The more steeply dipping delta front has relatively intense ice deformation and related gouging, whereas the more distal part of the delta is an area of relatively complex seasonal pack-ice movement. Because of the complexity of ice movement, both in space and time, the intervals of freezeup, winter, and breakup will be discussed separately.

Freezeup

In late October, as coastal temperatures drop below 0 °C, ice crystals typically begin to form and accumulate as new ice along the shore of Norton Sound. Bottomfast ice forms along the shallow margins of the delta (e.g., on intertidal mudflats and subaqueous levees); some of the smaller sub-ice channels begin to be covered by floating fast ice. Since the larger sub-ice channels that extend beyond the main distributaries are relatively deep and continue to maintain a channelized flow of water offshore, they are the last of the nearshore areas to freeze.

The shorefast ice continues to expand farther offshore in November, until the ice reaches a maximum width of from 15 to 30 km, approximately coincident with the limits of the sub-ice platform. Most of the shorefast ice is floating fast ice (Fig. 16-4), separated from the bottom fast ice by active tidal cracks that coincide approximately with the 1 m isobath. The inner zone of bottomfast ice is typically covered with a sheet of ice (*aufeis*) (Fig. 16-5), which forms by water flowing over bottomfast ice due to the tide- or storm-induced rise and fall of floating fast ice. The shorefast ice continues to expand seaward until it encounters mobile seasonal pack ice. At this time pressure ridges develop, become grounded, and a seaward-accreting *stamukhi* zone develops approximately coincident with the delta front in water depths of 5-15 m. The *stamukhi* zone is well developed by the beginning of December and, for the purpose of this report, marks the beginning of the winter period.

Winter

The winter phase of the ice-dominated regimen is characterized by the establishment of a relatively stable band of shorefast ice fringed by a complex zone of ice deformation features that form the *stamukhi* zone (as defined by Reimnitz et al. 1977). The patterns of ice movement seaward of the *stamukhi* zone are rather complex, reflecting both local and regional meteorologic events as well as the effect of bathymetry in deflecting and grounding ice floes.

Seasonal pack ice in Norton Sound is typically 0.7-1.2 m thick (Brower et al. 1977) and is largely *in situ*, having formed within a zone of ice divergence in the northeastern side of the sound. It usually flows to the west and southwest (Fig. 16-6) in response to the predominant northeasterly winds resulting from a relatively stable high-pressure system that develops during the winter months. Sluggish ice movement controlled by relatively weak oceanic currents occurs only during periods of weak winds. The southwest-flowing ice commonly converges against the shorefast ice fringing the Yukon Delta to form a broad *stamukhi* zone. The deformation of pack ice within the *stamukhi* zone causes pressure ridge raking which produces numerous parallel furrows as the ice keels plow through the bottom sediment (Reimnitz and Barnes 1974). As expected, the area of extensive ice gouging delineated by Thor et al. (1978) generally coincides with and parallels the *stamukhi* zones as delineated in this study. Ice seaward of the *stamukhi* zone is deflected to the west, where it leaves Norton Sound and joins the stream of rapidly moving Bering

Figure 16-4. Depositional environments of the modern lobe of the Yukon Delta (from Dupré and Thompson 1979).

Figure 16-5. Location of shear zones around the Yukon delta, as determined from LANDSAT imagery, 3/13/73 to 3/6/75. Hatching delineates areas of overflow icing (auteis). Dotted area delineates the 5-10 m bathymetric interval, which is approximately coincident with the delta front (see Fig. 16-4).

Sea pack ice. Pack ice rarely enters the sound from the west except during prolonged periods of strong westerly winds.

The pattern of ice movement to the west of

Norton Sound is relatively complex. In general the ice flows mainly to the south in response to the prevailing northerly winds. Significant reversals in the path of the flow can occur, however, as illustrated

Figure 16-6. Patterns of ice movement for 14-15 March 1974. Arrows indicate the direction and amount of ice movement in one day. Dashed lines within the shorefast ice to the north of the delta indicate inactive shear zones.

on the LANDSAT and weather data for 23-27 Feb. 1976 (Fig. 16-7). During this period, winds began to blow up to 25 knots from the south. As a result the southerly flow of ice ceased and northerly movement of ice as fast as 15 km/day was charted (Fig. 16-7). The rapid reversal was the result of the passage of two large low-pressure systems (Fig. 16-8A and B) and illustrates the extent to which ice movement is responsive to meteorologic events.

Shapiro and Burns (1975) noted that unusually strong northerly winds can result in a major ice deformation event in which masses of ice are deformed and funneled out of the Chukchi Sea and through the Bering Strait. Similar deformation features were noted on NOAA (VHRR) imagery for 13-15 March 1976. During this time, ice floes west of the Yukon Delta were moving as fast as 45 km/day (Fig. 16-9), presumably in response to a major ice deformation event similar to that described by Shapiro and Burns.

It is important to note that the zone of very rapid ice movement is restricted to a relatively narrow band approximately 70 km wide, which is referred to as the "racetrack." This zone is often recognized as a band of highly fractured nilas ice (Fig. 16-7) that presumably forms as the source of the pack ice (the Chukchi Sea) becomes temporarily plugged at the Bering Strait. The racetrack can be seen on LANDSAT imagery throughout much of the winter and recurs annually. The eastern boundary coincides approximately with the 20-m isobath, suggesting that this isobath may reflect the grounding of ice flows at the entrance to Norton Sound, and effectively deflects the ice to the south, parallel to the isobaths. There is often a zone of open water between the Bering Sea pack ice and the edge of the shorefast ice west of the Yukon Delta. This typically forms during periods of easterly offshore winds, but may be completely closed during periods of onshore winds. Some of it may, however, be caused by relatively deeper ice keels, grounded in water depths of 15-18 m, acting as an offshore ice barrier to some onshore movement of ice. This zone of open ice west of the delta is of particular interest to the natives in the area, as it greatly facilitates winter hunting.

Little, if any, sediment enters Norton Sound during the winter months, and yet the suspended sediment concentration measured beneath the ice in the west-central part of Norton Sound is essentially as high in winter as it is during much of the summer (personal communication, David Drake, U.S.G.S.). This suggests that a significant amount of sediment initially deposited during the summer months is resuspended during the winter and hence is available to be redistributed by sub-ice currents. The unusual off-

shore increase in sand off the Yukon Delta (Dupré and Thompson 1979) may come from such a process, but the exact mechanism(s) for such resuspension are unclear.

Breakup

Breakup along the coast is a relatively brief event that marks the transition between the ice-dominated and river-dominated regimens; the significance of breakup, however, far outweighs its brief duration. River breakup along the Yukon, as with most of the coastal rivers in northern Alaska, is marked by a tremendous increase in sediment and water discharge, resulting in ice jams, extensive inland flooding, and riverbank erosion.

As river discharge begins to increase, floating fast ice begins to lift, both in the river and along the coast. During this time, the sub-ice channels are especially well delineated by the floating fast ice. The bottomfast ice begins to be flooded by an over-ice flow (Fig. 16-10), which has been described for the North Slope by Reimnitz and Bruder (1972) and Walker (1974).

Some sediment is carried onto the ice, thereby effectively bypassing much of the inner sub-ice platform. Much of the sediment seems to remain in the sub-ice channels, which cross the sub-ice platform. Some of the sediment is probably deposited from suspension on subaqueous levees farther offshore; much of it, however, bypasses the sub-ice platform completely and is deposited on the delta front or prodelta. The floating ice that marks the sub-ice channels soon breaks up and is removed to sea. Much of the over-ice flow may drain through strudel holes (as defined by Reimnitz and Bruder 1972) or cause the bottomfast ice to melt in place. Large pieces of floating fast ice break off to be transported farther offshore. Grounded ice may remain in some shallow areas to the northwest of the delta; floating pack ice may remain trapped in the middle of Norton Sound because of the sluggish currents.

Figure 16-11 illustrates conditions that are typical during breakup. Floe movement is to the north as is common during late April and May when northerly winds die down and northward-flowing currents become more effective in transporting ice (cf. Muench and Ahlnäs (1976)). The floes were moving up to 20 km/day on 7 May 1974 (Fig. 16-10). A low-pressure system moved into the area on the 8th, however, bringing a temporary restoration of winter-phase northerlies. Temporary reversals of ice movement towards the south were also noted in late May 1974 by Muench and Ahlnäs (1976). These short-lived events are typically associated with

Figure 16-7. Patterns of ice movement for 23-27 February 1976. Arrows indicate the direction and amount of ice movement in one day. Dashed lines within the shorefast ice indicate inactive shear zones. Note the reversal in ice movement between 25-26 February and 26-27 February.

Figure 16-8. Weather conditions for the period 24 February to 27 February 1976 (from NOAA Surface Synoptic weather charts). Note the reversal of winds in the Bering Strait area due to the westward migration of low-pressure systems a and b.

the passage of a low-pressure system crossing the Bering Sea. Southerly winds typically follow, however, facilitating the breakup and removal of the shorefast ice. By early June the shorefast ice is

usually no longer present, although some areas of unconsolidated pack-ice floes may be present, particularly in the center of Norton Sound. At this time the distributary channels have been cleared of ice and

Figure 16-9. Patterns of ice movement for 13-15 March 1976. Arrows indicate direction and amount of ice movement in one day (up to 45 km/day).

Figure 16-10. Diagram of sediment dispersion patterns during spring breakup. Note the role of over-ice flow and sub-ice channels in providing mechanisms by which sediment may by pass the inner part of the sub-ice platform.

Figure 16-11. Patterns of ice movement for 7-9 May 1974. Arrow indicates direction and amount of ice movement in one day. The pack ice is much less consolidated than earlier in the year.

are introducing an apron of sediment-laden water over much of the prodelta region, marking the beginning of the river-dominated regimen.

SUMMARY

The patterns of ice formation, movement, and deformation in the Norton Sound region were studied with LANDSAT and NOAA satellite imagery for the years 1973-77. The results document not only the marked seasonality of marine processes throughout the year, but also the significant role of bathymetric and meteorologic conditions in controlling the patterns and rates of ice movement in the region. The results have been summarized in a map (Fig. 16-12) of generalized ice hazards similar in some respects to maps done for the entire Bering Sea by Stringer (1978). The following is a brief summary of the types of ice-related hazards that characterize each of the zones.

Zone Ia is a zone of shorefast ice that extends to the outer edge of the sub-ice platform of the Yukon Delta, approximately coincident with the 2-3 m isobath. Over-ice flow occurs throughout the winter in areas of bottomfast ice near the major distributaries (shown in hatched pattern). Sub-ice currents beneath the floating fast ice may result in some resuspension of sediments in the sub-ice channels and on the outer edge of the sub-ice platform. This is a relatively stable zone throughout the winter, but large sheets of ice may break off during spring breakup. *Zone Ib* is a slightly less stable area characterized by floating fast ice during most of the winter; the area can, however, be completely ice free under some conditions (e.g. 13-15 March 1976). *Zone Ic* is the zone of shorefast ice that fringes most of Norton Sound. It consists largely of floating fast ice, more variable in extent and less stable than ice in Zones Ia and Ib, because large sheets of ice may break off repeatedly throughout the winter.

Zone IIa is a broad, seaward-accreting stamukhi zone formed by the convergence and deformation of ice originating mainly in Norton Sound. The configuration of the outer margin of this zone seems to be controlled by Stuart Island to the east and a series of offshore shoals to the west, and coincides approximately with the 14 m isobath. This zone is characterized by extensive ice shearing and a relatively high intensity of ice gouging of the sea floor (Thor et al. 1978). *Zone IIb* is west of the delta in water 3-14 m deep. It is a relatively unstable area characterized by ice deformation and accretion to the shorefast ice (*Zone Ia*) during periods of westerly

onshore winds, or an offshore movement of ice and the development of a large polynya (open water area) during periods of easterly offshore winds. This zone is also characterized by moderately high density of ice gouging.

Zone III is an area of seasonal pack ice formed mainly *in situ*, within Norton Sound. The ice typically moves south and west in response to the dominant northeasterly winds throughout the winter, but it may drift slowly in response to oceanic currents during periods of low winds. The southern portion of this zone is characterized by widespread shearing of ice and coincides approximately with the area of very high density of ice gouging delineated by Thor et al. (1978). The western boundary coincides approximately with the 20-m isobath, separating pack ice formed in Norton Sound from the thicker pack ice formed farther to the north. Bering and Chukchi pack ice enter the sound only rarely when especially strong northwesterly winds blow.

Zone IV consists of seasonal pack ice formed in the northern Bering and Chukchi Seas. The ice typically moves southward in response to northerly winds for most of the winter. Short periods of northerly ice movement can, however, occur during the passage of low-pressure systems. It is typical for the ice to begin to move consistently to the north in late April or early May. *Zone IVa* is the "racetrack," characterized by intervals of extremely rapid southerly movement of pack ice (up to 45 km/day) following major ice-deformation events north of the Bering Strait (described by Shapiro and Burns 1975). This zone is characterized by highly fractured nilas ice during periods of relative quiescence. The eastern margin of this zone is approximately coincident with the 22 m isobath. The more variable western margin appears to be controlled by the geometry of ice piling up on the northern side of St. Lawrence Island. The rapid ice movement in this zone, combined with the lack of ice gouging (Thor et al. 1978) suggests that this is a zone where ice is unlikely to affect the bottom. *Zone IVb*, in water depths of 20-22 m, is characterized by less rapid ice movement than in the racetrack. Some grounded ice may occur in this zone, particularly in the area of shoals southwest of the delta. *Zone IVc*, in water depths of 14-20 m, is characterized by open water during periods of easterly winds and onshore moving pack ice during periods of westerly winds. This zone rarely has a stamukhi zone accreted to the shorefast ice, although some grounded ice and ice gouging will occur. *Zone IVd* is similar to zone IVb, and was not studied in detail.

Figure 16-12. Zonation of ice hazards in the Yukon Delta/Norton Sound region based mainly on LANDSAT and NOAA satellite imagery, supplemented information on ice gouging from Thor and Nelson (1979).

Zones Va and Vb are zones of ice divergence formed by persistent offshore winds (cf. Muench and Ahlnäs 1976). They are typically areas of open water where ice is actively forming for most of the winter.

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Ice Gouging on the Subarctic Bering Shelf

17

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ABSTRACT

Ice striking the sea floor gouges surficial sediment of the shallow Bering epicontinental shelf of Alaska. Two types of ice gouge have been recognized: the single gouge, a single furrow, and multiple gouges or raking, a wide zone of numerous, subparallel furrows. Single gouges, the most common type, are cut by single-keeled pieces of thick ice, whereas multiple gouges are formed by multikeeled, thick, pressure-ridge ice. Gouges occur in water depths of 30 m or less, but are most dense in water 10 to 20 m deep. Although some gouge incisions are as deep as 1 m, most are 0.5 m or less. Ice gouges trend parallel to pack-ice movement, which in turn is generally parallel to isobaths and coastline configuration. Mean gouge trend in Norton Sound is west-east, in the northeastern Bering Sea north-south.

The annual ice cover in this subarctic setting is thin (less than 2 m). Ice thick enough to gouge the substrate forms in compression and in shear zones; there moving pack ice collides with and piles up against other pack ice or stationary shorefast ice to develop pressure ridges. Southward-moving pack ice in the northeastern Bering Sea and westward-moving pack ice in Norton Sound converge with, and shear past, a shorefast ice zone 10-30 km wide that covers the shallow water offshore of the Yukon Delta. The intensity of ice deformation in this zone causes the highest gouge density in the study area. In contrast, northeastern Norton Sound is an area of ice divergence and only minimal ice gouging. The rest of Norton Sound and northeastern Bering Sea is either in ice-divergence areas or else water depths are too great for ice to touch bottom, and thus ice gouge density is low. Gouging is extremely rare inshore of the shear zone, because shorefast ice is relatively static and protects inshore areas from the dynamics of the shear or compression zone and consequent ice gouging.

INTRODUCTION

Development of natural resources in northern latitudes has led to increased research on the effects of ice on shelf sediment in arctic regions like the Beaufort Sea (Reed and Sater 1974, Reimnitz et al. 1973, Reimnitz et al. 1977, Barnes et al. 1978). Until recently, however, research on ice gouging had not been done in subarctic regions like the Bering Sea. A variety of gouge features are found in many areas of the northeastern Bering Sea, even though ice

conditions there are not so severe as in high-latitude arctic regions. Ice gouging into the sea floor is a potential hazard to future resource development and such sea-floor installations as pipelines and wellheads.

This chapter discusses general ice conditions and ice movement in the northeastern Bering Sea, the effect of ice as an erosional and depositional agent that influences the geomorphology and depositional history of the shallow subarctic Bering Sea shelf, and ice gouging as a potential hazard to resource development in and around Norton Basin. The terminology used is adopted from Barnes et al. (1978), particularly the word "gouge" to describe the feature and the process of ice interacting with the sea floor.

Geographic setting

The floor of the northeastern Bering Sea is a broad, shallow epicontinental shelf (Figs. 17-1 and 17-2). Water depths in Chirikov Basin range from 20 m on the eastern side to 50 m in the central part. The shelf is generally flat and featureless except for a prominent series of ridges and swales that subparallel the coastline off Port Clarence. A large, elongate marine reentrant forms Norton Sound, bounded on the north by Seward Peninsula, on the east by the Alaskan mainland, and on the south by the Yukon Delta. Except in a broad trough in the northern part of the sound, where depths are as great as 27 m, water depths in Norton Sound range from 10 to 20 m. The offshore part of the Yukon Delta is a zone of extensive shoals covering about 8,000 km² (Fig. 17-2). Water depths 10-30 km offshore do not exceed 3 m; beyond that zone there is a gentle break in slope and the depth increases to 10 m as far as 50-70 km from shore. The substrate of the Yukon prodelta, derived from the Yukon River, consists of

Figure 17-1. Index map and chart of high-resolution geophysical and side-scan sonar tracklines covered by the R/V Sea Sounder and R/V Karluk in northeastern Bering Sea during 1976, 1977, and 1978.

Figure 17-2. Northeastern Bering Sea and southern Chukchi Sea, showing water circulation and bathymetry. Compilation sources include Goodman et al. (1942), Fleming and Heggarty (1966), Husby (1969, 1971), McManus and Smyth (1970), Nelson and Hopkins (1972), Pratt and Walton (1974), and Coachman et al. (1975). Drift directions of pack ice in northern Bering Sea adapted from Muench and Ahlén (1976) and Dupré (1978).

coarse silt to very fine sand, whereas sediment in Chirikov Basin consists mostly of glacial gravel and transgressive fine sand (Nelson and Hopkins 1972, McManus et al. 1977).

Ice conditions and movement

Ice overlies the northern Bering Sea annually from November through June (Muench and Ahlnäs 1976, Shapiro and Burns 1975). Depending on the severity of the winter, multiyear ice may migrate into the Bering Sea from the southern Chukchi Sea. The keel depth of 90 percent of the pack ice (any free-floating ice regardless of origin) is less than 1 m, although depths to 20 m have been reported (Arctic Research Laboratory 1973).

Ice in open sea pans in Norton Sound is 0.7-1.2 m thick (Brower et al. 1977), but can get as thick as 2 m (C. H. Pease 1979, personal communication). Shorefast ice (ice anchored to the land) extends seaward to about the 10 m isobath and is best developed in the southern part of Norton Sound, around the Yukon Delta (R. E. Hunter, personal communication, 1976; Dupré 1977; Stringer et al. 1977; Ray and Dupré, this volume) (Fig. 17-2).

Analysis of LANDSAT photographs (Ray and Dupré, this volume; Stringer et al. 1977; Muench and Ahlnäs 1976; Shapiro and Burns 1975) has contributed to a preliminary understanding of ice dynamics in the Bering Sea. Pack ice in the northern Bering Sea originates as (1) *in-situ* northeastern Bering Sea ice and (2) advected Chukchi Sea ice. Chukchi Sea ice can move through the Bering Strait and into the northern Bering Sea during episodes of rapid deformation and subsequent rapid southerly movement of pack ice caused by episodes of strong northerly winds (Shapiro and Burns 1975).

Ice movement in the northeastern Bering Sea is controlled by the interplay of (1) prevailing winter northeasterly geostrophic wind (Muench and Ahlnäs 1976), (2) erratic onshore wind (NOAA 1974), (3) northward-flowing water current on the eastern side of the Bering Sea (Coachman et al. 1976) (Fig. 17-2), and (4) a counterclockwise current gyre in Norton Sound (Nelson and Creager 1977) (Fig. 17-2). Late winter and early spring winds tend to push ice generally southward in the northeastern Bering Sea, whereas waning late spring winds allow pack ice to be increasingly influenced by the northward-flowing water currents (Fig. 17-2).

In Norton Sound the dominant direction of ice movement is southwestward out of the sound. This drift creates a zone of divergence in the northeastern part of the sound and a zone of convergence in the southwestern or Yukon prodelta area of the sound

(Ray and Dupré, this volume; Stringer et al. 1977) (Fig. 17-2). Periodic changes in wind and water current tend to move ice in and out of the sound, thereby making it possible for Bering Sea ice, or even advected Chukchi Sea ice, to work its way into the sound.

Zones of convergence can be zones of pressure-ridge or shear-ridge formation characterized by colliding, piling up, and deforming of the edges of fast ice and of pack ice (Reimnitz and Barnes 1974). The best-developed pressure ridges in the northeastern Bering Sea form around the Yukon Delta, where Bering Sea pack ice on the western prodelta and Norton Sound pack ice on the northern prodelta collide with the Yukon Delta fast ice (Ray and Dupré, this volume; Stringer et al. 1977).

Methods

Data for this study were gathered by the U.S. Geological Survey during September 1976, July 1977, and September 1978 aboard R/V *Sea Sounder* and during June and July 1978 aboard R/V *Karluk*. Approximately 5,100 km of side-scan sonar trackline was obtained (Fig. 17-1). Normally, seismic units with energy sources of 200 kHz, 12 kHz, 7 kHz, and 2 kHz were run simultaneously with side scan for additional bottom and subbottom information. The 6-m keel depth of the R/V *Sea Sounder* limited ship operations to water deeper than 8 m, whereas the shallow draft of the R/V *Karluk* (1 m) allowed surveying in nearshore areas and in the shallow waters off the Yukon Delta. Geophysical and navigational operations are described in Thor and Nelson (1978).

An EG&G¹ side-scan sonar system, consisting of a dual-channel graphic recorder and a towed transducer fish, was used to survey the sea floor. Side-scan sonar, an alternative method to conventional vertical echo sounding, employs a 105-kHz acoustic beam with an axis slightly below horizontal. This acoustic beam can resolve topographic irregularities and objects on the sea floor with as little as 10 cm of relief. Reflected echoes are recorded graphically in a form that approaches a plan view map. Discussions on theoretical and practical aspects of side-scan operation and interpretation can be found in Belderson et al. (1972) and Flemming (1976). Normally the side-scan was operated at 100-m sweep (the scan range on either side of the ship), but at times the 50-m sweep was used to help resolve details of the gouging. In addition, a 200-kHz high-resolution

¹ Any use of trade names and trademarks in this publication is for descriptive purposes only and does not constitute endorsement by the U.S. Geological Survey.

fathometer was operated to measure the incision depth of ice gouges (Fig. 17-3). Vertical relief of gouges on the fathometer record or on the horizon line of sonographs is generally masked by the recording of sea swell or ship's motion on the chart paper.

Gouge data were collected from the sonographs by counting the number, measuring the trend, and noting the location of all gouges seen on the records. Distortion of sea floor features on the sonograph occurs parallel to the line of travel because of the difference in ship's speed and the recorder's paper-advance speed. To obtain absolute compass trend of gouges, a distortion ellipse protractor, which corrects for the apparent angle produced by ship paper speed, was used to measure gouge angle with relation to ship's track. This information was then normalized at 10-km intervals. Normalization entailed two procedures: correcting the number of observed gouges and averaging observed gouge trends. The number of observed gouges per 10-km interval was multiplied by $1/\sin$ (where angle equals the angle between ship's course and gouge trend) to correct for the fact that ship's course usually was not normal to the gouge trend. Any angle other than 90° between ship's course and gouge trend will give a false picture of gouge density (Barnes et al. 1978). Averaging observed gouge trends involved graphing the measured trends for the 10-km intervals and noting the average dominant and subordinate trend or trends. Each average trend per 10-km interval was then plotted on the base map to define areas of similar gouge trend.

GEOMETRY AND TYPE OF ICE GOUGING

Two basic types of ice gouge have been recognized on the sea floor of the northeastern Bering Sea: single gouges and multiple gouges or raking. A single gouge, the dominant type of ice-produced mark on the Bering Sea floor, is a groove produced by a single ice keel plowing through the surficial sediment (Figs. 17-3A, 17-3B, 17-4A, 17-4B, and 17-4C) (Reimnitz et al. 1973, Reimnitz and Barnes 1974). Single gouges are ubiquitous throughout Norton Sound, but the highest density occurs around the prodelta of the Yukon River (Fig. 17-5).

Figure 17-3. A. Solitary gouge on a sonograph. B. 200 kHz fathometer profile and diagrammatic representation of gouge shown in A. Features of gouge include (a) incision depth as measured from gouge bottom to a horizontal line projected across sediment surface, (b) height of sediment mounded on the gouge end, (c) width of incision, (d) width of disruption zone caused by the gouging process. C. Box core slab showing subsurface (11-18 cm interval) disruption possibly caused by a past gouge event.

Figure 17-4. Sonographs showing ice gouges of the northeastern Bering Sea. A and B. Solitary gouges in sand-wave and ripple fields. C, D, and E. Solitary gouges. Example E shows depth of incision on the sonograph horizon line. F and G. Examples of pressure ridge raking. Example G shows depth of incision on the sonograph horizon line. H. Example of depressions associated with ice gouging.

Figure 17-5. Rose diagrams representing trend and density of gouges. Division into areas I-V based on zones of similar trending gouges. Zone of shorefast ice based on evaluation of Landsat imagery (Dupré 1977, 1978; R. E. Hunter, pers. comm. 1977).

Single gouge widths range from 5 to 60 m; a width of 15-25 m is most common. Gouge patterns range from straight through sinuous to sharp-angled turns (Fig. 17-4). Incision depths of gouges, as measured on the sea-floor profile of sonographs (Fig. 17-4E) and on the 200-kHz fathometer record (Fig. 17-3B), can be as much as 1 m, but most gouges range in depth from 0.25 to 0.5 m or less. These figures may be conservative because of the geometric relation between the narrow width of the gouge and the spread of the acoustic cone of the fathometer transducer (Reimnitz et al. 1977). The original incision depth is impossible to determine unless the gouge is

seen as the keel plows the bottom, because afterward the gouge will be filled in.

Multiple gouges, or raking (Figs. 17-4F and 17-4G) are produced when multi-keeled floes (such as pressure ridges) plow or rake the bottom sediment, creating many parallel furrows (Reimnitz et al. 1973, Reimnitz and Barnes 1974). Unlike single gouges, raking is not ubiquitous, but in the Yukon prodelta area the raking process is more prevalent than single gouging. Zones of raking are 50-100 m to several km wide. The deepest incisions caused by raking observed on the records are about 1 m; but raking, like

single gouges, usually produces incisions less than 0.25-0.5 m deep.

TRENDS AND DISTRIBUTION OF GOUGES

Analysis of the trend and distribution of gouges allows us to recognize five areas of gouging with similar trends (areas I-V) and two large areas almost devoid of gouges (VI and shorefast ice zone) (Fig. 17-5). Absolute direction of ice movement cannot be predicted because criteria needed to make certain distinctions, such as gouge terminations, were not seen on the sonographs.

In areas I and II (Fig. 17-5), the dominant trend of gouges is distinctly subparallel to isobaths and the coastline. There is more data scatter in areas III, IV, and V, but gouges again are generally parallel to isobaths and the coastline. The greatest data scatter is seen in area V, but this may reflect the irregular bathymetry of ridge and swale topography off Port Clarence. Except for a couple of gouges off the northwestern end of St. Lawrence Island, area VI is devoid of ice gouges.

Density of ice gouges is as much as 25 times higher around the Yukon Delta, where the water is 10-20 m deep, than in other areas of the northeastern Bering Sea (Table 17-1 and Fig. 17-5, areas I and II). Not coincidentally, the Yukon prodelta is the largest expanse of shallow water in the study region. Here the density of ice gouges can be as high as 75 gouges/km². Density of ice gouging is 60 times higher in water 10-20 m deep than in water 5-10 m or 20-39 m deep (Table 17-2). Gouging has not been seen in water shallower than 5 m or deeper than 30 m.

TABLE 17-1

Gouge density by area

Area	km ²	Trackline km	Total number of gouges	Average density (gouges/km ² *)
I	5,500	530	1,684	3.18
II	8,000	1,005	5,080	5.05
III	9,500	1,100	917	0.83
IV	15,500	400	993	2.48
V	7,900	1,120	216	0.19
VI	50,400	766	4	0.03

* Assuming that 1 km trackline of side-scan sonar is representative of 1 km².

TABLE 17-2

Gouge density by water depth interval

Depth interval(m)	km ²	Trackline km	Total number of gouges	Average density (gouges/km ² *)
0-10	16,500	480	147	0.31
10-20	24,600	2,100	8,593	4.09
20-30	32,700	1,300	143	0.11
30-40	26,000	750	0	0
40-50	12,600	450	0	0
>50	5,400	170	0	0

* Assuming that 1 km trackline of side-scan sonar is representative of 1 km².

GEOLOGICAL SIGNIFICANCE

Trend and density of gouges

The interplay of geomorphology, water depth, oceanic conditions, and location of compression or of shear zones (Fig. 17-2) determines the pattern of ice gouging in the northern Bering Sea (Figs. 17-5 and 17-6). The orientation of ice gouges depends on the direction of ice drift under the influence of wind and water current. The dominant trend of ice gouges, therefore, in Norton Sound is east-west and in the Bering Sea north-south (Figs. 17-5 and 17-6).

Land promontories like the Yukon Delta tend to block ice movement and to cause compression and shear zones to form. Ice ridges around the Yukon Delta formed by the collision and shearing of moving pack ice with stationary shorefast ice account for the high density of ice gouges in areas I and II (Fig. 17-5). Areas within the zone of shorefast ice, like the large area around the Yukon Delta (Fig. 17-5), are devoid of gouges. Only the edge of the shorefast ice is deformed by the pack ice, and subsequent deformation occurs continually seaward through a process of migration of the compression/shear zone through time (Dupré 1978). Areas III and IV are characterized by low density of ice gouges (Fig. 17-5). Gouging in areas III and IV is the product of ridges formed in an ice-divergence zone by intercollisions of pack ice. The density of ice gouges in area V is low because this area is not in a convergence zone and at most places water depth exceeds normal ice-keel depths. Area VI seems not to have any ice gouging because water depths (Fig. 17-2) exceed normal ice-keel depths (Fig. 17-5).

Age of ice gouges

Although no specific studies were made to determine the age and longevity of gouges, the gouges

Figure 17-6. Summary of ice gouging: density, shorefast ice limits, and ice movements in northeastern Bering Sea.

seem to be modern ephemeral phenomena that recur annually. West of Port Clarence and in the nearshore area of Nome, ice gouges cut through ripple and sand-wave fields that are in dynamic equilibrium with present wave or current motion (Nelson et al. 1978, Hunter and Thor 1979) (Figs. 17-4A and B). Here the juxtaposition of old gouges, highly modified by ripples or sand waves, with new gouges suggests that gouges are being formed each winter.

A number of geologic processes act to destroy gouges rapidly once they have formed. The initial smoothing of ice gouges can be enhanced by: (1) the saturated, silty substrate that tends to seek a minimum relief equilibrium with sides of the gouge flowing or slumping toward the center of the gouge,

and (2) the constant oscillatory pounding of wave motion on the sea floor that causes shear failure in the soft sediment (Henkel 1970), making gouge sides collapse toward the center. The dish-shaped profiles of most gouges (Figs. 17-4E and 4G) indicate that these processes normally occur.

Repeated surveys of ice gouges in water less than 20 m deep in the Beaufort Sea have shown that gouges are frequently smoothed over completely in one season (Barnes and Reimnitz 1979). In the Bering Sea, the ice-free season is three to four months longer than in the Beaufort Sea, allowing more time for the considerably stronger open-water wave and current regimes of the Bering Sea to destroy gouges. In Norton Sound, storm waves and currents caused

by the advance and retreat of storm-surge water, in addition to normal tidal and geostrophic currents, resuspend and transport large quantities of surficial sediment (Cacchione and Drake 1978, Nelson and Creager 1977). Destruction of gouges is augmented by biological reworking of surficial sediment, an active process in Norton Sound (Nelson et al., volume 2). In summary, gouges tend to be either eroded or buried because they are not in equilibrium with the dynamic physical processes on the sea floor. This reinforces the hypothesis that gouges in the Bering Sea are present-day phenomena involving development of some new gouges each ice season.

Ice-sediment interaction

Ice acts as both an erosional and a depositional agent; it gouges, mixes, and deforms the substrate, and promotes current scour. Ice partially controls the geomorphology of the Yukon Delta (Dupr and Thompson 1979).

Sediment mixing and deformation of the substrate are important processes in densely gouged areas such as the Yukon prodelta, where pressure-ridge raking can gouge 1 m into the sediment. One event of pressure-ridge raking can affect several square kilometers of sea floor,² and mix or disrupt several million cubic meters of sediment. A zone of deformed sediment in box core No. 48 (11-18 cm interval, Fig. 17-3C) may represent an ice-gouge event.

The sharpness of gouge morphology is highly dependent on the type of substrate being gouged. The sediment of the Yukon prodelta is a moderately cohesive sandy silt that will hold a shape better than the coarser-grained sediment of central Norton Sound or offshore from Port Clarence (Clukey et al. 1978, Nelson and Hopkins 1972, McManus et al. 1977). The gouge shown in Fig. 17-5A and some gouges shown in Fig. 17-4A are examples of forms with sharp relief in a competent substrate. Groups shown in Fig. 17-4A are smoother in form because they cut into a cohesionless sand substrate in the Port Clarence area.

Prominent broad (50-150 m wide), shallow (0.6-0.8 m deep) depressions on the western Yukon prodelta are associated with areas of intense ice gouging and strong bottom currents (Larsen et al. 1979). Topographic disruption by ice gouges in these areas apparently causes flow separation in the strong currents, thereby initiating scour depression for

extensive distances downstream. Consequently, large regions of scour may continue to expand away from intensely gouged areas (Fig. 17-4H).

The extensive depositional sand shoals of the Yukon Delta front coincide with the seaward extent of shorefast ice, stamukhi (grounded pressure ridges), and zones of dense ice gouging (Figs. 17-2 and 17-6). Reimnitz and Barnes (1974) have noted this relation in the Colville Delta area of the Beaufort Sea. They postulate that pressure ridges and stamukhi act as sediment traps or dams, channelize winter currents, or bulldoze sediment to form shoals. Thus, a cycle is formed in the sense that shoal areas determine the extent of shorefast ice and the location of a shear zone and pressure ridges, which in turn cause shoals to develop. Dupr (1978) hypothesizes that the geomorphology of both onshore and offshore parts of the Yukon Delta is controlled by ice.

RESOURCE DEVELOPMENT: POTENTIAL HAZARDS

To summarize, gouges are ubiquitous throughout the northeastern Bering Sea in water depths of 5-30 m. Ice-gouge density varies from rare to sparse in the northeastern Bering sea and northern Norton Sound; the maximum density is around the Yukon Delta (Fig. 17-6). The depth of ice gouges is fairly uniform throughout the northeastern Bering Sea and seems to be independent of gouge density. Although the maximum observed ice-gouge depth is about 1 m and maximum observed current scour about 1 m, the combination of these forces could affect the bottom to depths of several meters, thus presenting some design problems and potential hazards to installations in or on the sea floor. Pipelines and cables should be buried below the combined effective depth of ice gouging and current scour, plus a safety factor.

Special studies of nearshore areas off Nome and Port Clarence were conducted because both are potential centers for commercial development and activity. Nome, already a well-established small city, is the focal point for barge traffic in the northern Bering Sea. Port Clarence, the only natural harbor in the northern Bering Sea, has high potential for development as a site for future shipping activity.

Offshore Nome, being an area of ice divergence, is not heavily gouged. Although some gouges were found offshore, none were in water shallower than 8 m, and several are probably not related to ice. They are narrower (less than 1 m) than typical ice gouges (more than 5 m wide) and may have been produced by anchor, anchor chain, or cable drag from the tugs and barges that frequent the port of Nome.

² Area of gouging times depth of gouging. Ex.—2,000 m (length of gouged zone) × 1,000 m (width of gouged zone) × 0.5 m (depth of gouge) = 1,000,000 m³.

Several gouges were found near Port Clarence at the northern end of the Port Clarence spit and on the northern side of Port Clarence inside the tidal inlet, but none occurred in water less than 8 m deep.

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The Role of the Kaltag and Kobuk Faults in the Tectonic Evolution of the Bering Strait Region

18

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ABSTRACT

The generally latitudinal trend of the major basins and mountain belts in northwestern Alaska and northeastern Siberia places severe constraints on palinspastic reconstructions of the tectonic features of this region. The essential continuity of these linear trends since Precambrian time precludes the possibility that there has been any significant north-south relative movement between Alaska and Siberia.

The geological similarity between the Seward and Chukotsk Peninsulas suggests that they probably acted as a single unit, bounded on the south by the Kaltag fault and on the north by the Kobuk fault. This tectonic block moved eastward more than 100 km during the Tertiary, partly as a result of sea-floor spreading in the North Atlantic and Arctic Oceans. An important assumption in this model is that the Kobuk Trench represents a major left-lateral transcurrent fault.

INTRODUCTION

In any attempt at palinspastic reconstruction or mobilistic modeling of the major tectonic features of northwestern Alaska, serious consideration must be given to the long-term similarities in the geologic histories of northern Alaska and northeastern Siberia. The generally latitudinal distribution of the major mountain ranges and geosynclinal basins has led several writers (Gates and Gryc 1963; Smirnov 1968; Churkin 1969, 1970, 1972, 1973; Lathram 1973; and Meyerhoff 1973) to point out that these features have had an essential continuity since Precambrian time, and that the close relationship of adjacent elements precludes the possibility that there have been any major relative movements between Alaska and Siberia. This view holds that the 1,350-km-wide Bering-Chukchi Shelf is an integral part of both North America and Eurasia which has accreted between the Canadian and Siberian shields since Proterozoic time. The continuity of many of the major geologic fea-

tures certainly does seem to preclude the possibility that there has been any major movement at right angles to their roughly east-west trends. However, there is abundant evidence of significant amounts of late Mesozoic and Tertiary thrusting and transcurrent faulting that imply differential east-west tectonic transport; this type of tectonic activity would maintain an apparent continuity in the older geologic trends.

GEOLOGICAL SETTING

Figs. 18-1 and 18-2 show the geography and generalized geological and structural elements of the Arctic area of western Alaska and eastern Siberia, respectively. Rocks ranging in age from Precambrian to Pleistocene occur in the several mountain belts and extensive coastal plain zones of both Alaska and northeastern Siberia. Two sedimentary basins occupy the continental shelf areas north and south of Bering Strait. These large and complex structural troughs contain rock sequences possibly as old as Late Cretaceous.

Quaternary nonmarine sediments forming the lowland 75 km wide on the northern side of the Chukotsk Peninsula rest on Paleozoic sedimentary, metamorphic, and igneous rocks and early Tertiary intrusives (Nalivkin 1960, Suslov 1961). These rocks also form the main spine of the Chukotsk Range. Mountains of the Tenilny Range consist of Precambrian and Paleozoic sedimentary and metamorphic rocks, with some Mesozoic intrusive and Cenozoic extrusive rocks (Nalivkin 1960, Markov and Tkachenko 1961).

Figure 18-1. Geographic features of northwestern Alaska and northeastern Siberia.

The geology of Seward Peninsula is strikingly similar to that of the Chukotsk Peninsula. Paleozoic and some Precambrian sedimentary and metamorphic rocks underlie most of the peninsula, with extensive occurrences of Tertiary extrusives near Cape Espenberg and in the eastern part of the peninsula. The rocks of Seward Peninsula are cut by numerous zones of thrust faults, along which the movement has been eastward (Sainsbury 1969).

From Cape Krusenstern northwest to Cape Lisburne, the coastline consists for the most part of cliffs formed by rocks of the Brooks Range, the De Long Mountains, and the Lisburne Hills. The Brooks Range is formed by a belt of predominantly Paleozoic rocks, and Early Cretaceous orogenic rocks have been thrust northward, forming tectonic sheets of Jurassic mafic and ultramafic rocks and Devonian and Mississippian carbonate rocks unlike the Devonian schists and Mississippian rocks which they override. North of this trend is a narrow zone of chaotically deformed lower Mesozoic and upper Paleozoic rocks that has been called the Disturbed Belt (Tailleur and Brosgé 1970, Brosgé and Tailleur 1970).

The southern margin of the Brooks Range (Fig. 18-2) is formed by the Kobuk Trench (Grantz 1966). South of this feature, the rocks of the Yukon-Koyukuk basin consist of strongly deformed Permian to Jurassic volcanic and Cretaceous sedimentary and volcanic rocks (Payne 1955, Patton 1973). The southern part of the Yukon-Koyukuk basin is cut by the Kaltag fault (Fig. 18-2), a major northeast-trending trans-

current fault showing right-lateral displacement (Patton and Hoare 1968).

The Lisburne Hills, extending from Cape Lisburne on the northwest to Cape Thompson in the southeast, show structures similar to those of the Brooks Range; here the Paleozoic rocks have been thrust eastward in imbricate thrust sheets over the Cretaceous rocks of the Colville Geosyncline (Campbell 1967, Martin 1970).

The marine geology of Norton basin (Fig. 18-2) has been discussed by Moore (1964), Scholl and Hopkins (1969), Grim and McManus (1970), Tagg and Greene (1973), Nelson et al. (1974), Holmes and Cline (1978), Holmes and Fisher (1979), and Fisher et al. (1979). The basin consists of a trough 5 km deep (Holmes and Fisher 1979) that may be filled with Upper Cretaceous and Paleogene nonmarine rocks and Neogene marine deposits. The basin is underlain by an acoustic basement characterized by velocities of 4.5 to 6.9 km/sec, probably consisting of metamorphosed Paleozoic and upper Mesozoic rocks which have been locally intruded by upper Mesozoic plutons similar to those occurring on St. Lawrence Island and which also form the several islands in northern Norton basin (Holmes and Fisher 1979).

North of Bering Strait another sedimentary basin underlies much of the southern Chukchi Sea (Fig. 18-2). Hope basin (Grantz et al. 1970, 1975; Eittreim et al. 1978), is similar in many respects to Norton basin. It contains approximately 3-3.5 km of sedimentary fill consisting of possible Upper Cretaceous and Paleogene coal-bearing nonmarine deposits and Neogene marine sediments.

The acoustic basement beneath this sedimentary basin is characterized by compressional velocities ranging from 4.3 to 5.2 km/sec (Grantz et al. 1975), indicating that basement may consist of rocks similar to the Mesozoic and Paleozoic carbonate and clastic rocks and Precambrian metamorphic rocks which have been mapped onshore around the margins of the basin.

The northeastern and northern margin of the basin is formed by Herald and Wrangel arches (Grantz et al. 1975), two uplifted areas where the Mesozoic and Paleozoic basement rocks have been thrust northward and northeastward over younger Cretaceous rocks of the Colville Geosyncline. These arches form a structural and stratigraphic continuation of the Lisburne Hills uplift. Grantz et al. (1975) have shown that the Herald Fault Zone along the northeast side of Herald Arch is an offshore extension of the thrust front along the Lisburne Hills (Martin 1970).

Northwest of the Herald Fault Zone, Grantz et al. (1970, 1975) were able to trace the westward offshore extension of the Colville Geosyncline structures

Figure 18-2. General geological and structural elements of northwestern Alaska and northeastern Siberia. Compiled from Nalivkin (1960), King (1969), Sainsbury (1972), Patton (1973), Beikman (1978).

and show that they appear to be truncated by the faults and folds along the front of the Herald Fault Zone.

THE KOBUK FAULT

The Kobuk Trench, or Trough, has been described by Grantz (1966) and Patton (1973) as a possible strike-slip fault which extends eastward for over 480 km across northern Alaska to the point where it intercepts the northeast-trending Kaltag fault (Fig. 18-2). The southward deflection of major structural trends on the north side of the Kobuk Trench was thought by Patton (1973) to represent large-scale drag folding due to major left-lateral movement. ERTS imagery studies have revealed a series of cross faults which could be tensional features indicating left-lateral displacement along the Kobuk strike-slip fault (Lathram 1972, 1973; Baker 1974). In the valley of the Kobuk River the deformational zone is approximately 20 km wide. The age of faulting is thought to be pre-middle Tertiary, but some local rupturing of Pleistocene drift (Patton 1973) indicates that movement has probably continued through Tertiary and into Pleistocene time.

Grantz et al. (1975), Holmes et al. (1968a), Holmes (1975), and Eittreim et al. (1978) show a prominent west-trending ridge formed by the acoustic basement surface in the southeastern part of Hope basin (Fig. 18-2). The ridge appears to be approximately 150-200 km long and 20-22 km wide with a relief of 250-275 m. Several faults can be seen cutting the basin fill along the flanks of the ridge (Grantz et al. 1975) and the upper Neogene (?) unit in the basin has been uplifted and eroded over the crest of the ridge (Holmes 1975), forming a pronounced angular unconformity with the base of the Holocene section (Holmes et al. 1968b).

The basement uplift and associated west-trending structural features are situated in such a way as to appear to be the offshore extension of the Kobuk Fault Zone. Marine gravity measurements (Ruppel and McHendrie 1976) also indicate a correlation between the Kobuk Trench and this offshore feature.

THE KALTAG FAULT

The Kaltag Fault (Fig. 18-2) is another major transcurrent fault which trends in a northeastern direction for over 440 km across north-central Alaska from Norton Sound. Scholl et al. (1970) have traced the offshore continuation of the Kaltag fault for over 320 km to the southwest beneath St. Matthew and Hall basins. Patton and Hoare (1968) mapped several pro-

vince boundaries and geologic trends which have been displaced up to 130 km right-laterally since Cretaceous time; bending of the structural grain near the fault was ascribed by Patton (1973) to large-scale drag folding.

Marine seismic refraction data suggest that the oldest sedimentary unit in Norton basin may consist of Upper Cretaceous rocks similar to those in the Yukon-Koyukuk basin (Holmes and Fisher 1979). If this interpretation is correct, it could indicate that movement along the Kaltag fault began in late Cretaceous time. Deformation of sedimentary units in St. Matthew basin (Fig. 18-2) suggests movement along the fault at least by early-middle Tertiary time (Scholl et al. 1970). Grantz (1966) and Patton and Hoare (1968) describe offset drainage patterns and displaced strata along the onshore segment of the fault, indicating right-lateral movement along the fault of as much as 2.4 km during the Holocene.

MODEL FOR MAJOR HORIZONTAL DISLOCATIONS

A possible evolutionary model may explain the present configuration and relationship of the major tectonic features observed in the Bering Strait region, including western Alaska and eastern Siberia. Elements of the late Mesozoic and Cenozoic geologic history are presented in outline form to weave together a coordinated account of a complex sequence of events which relates the major horizontal movements (with the exception of the longitudinal faulting of the Brooks Range and De Long Mountains) to a relative eastward movement of Siberia toward Alaska. We do not intend to imply that a crustal plate boundary exists between these two continental areas, even though others have proposed such a feature (Le Pichon 1968, Hamilton 1969).

Two basic assumptions are made: (1) Seward Peninsula represents a direct extension of the geologic trends of the Chukotsk Peninsula rather than of the Alaskan Yukon-Koyukuk basin (Sachs and Strelkov 1961, Sainsbury 1972); (2) the Kobuk Trench does, in fact, represent a major left-lateral transcurrent fault (Lathram 1973, Baker 1974), which has been active at least since early Tertiary time (Patton 1973). A corollary assumption is that the east-west ridge/trough structure in southeastern Hope basin (Fig. 18-2) is the offshore extension of the Kobuk fault.

The central feature of the model is the late Cretaceous development of a single continuous arcuate belt of deformation and thrusting extending from eastern Seward Peninsula northwest across the southern Chukchi Sea to Wrangel Island. The contemporaneity

of various segments of this arc has already been pointed out by Grantz et al. (1975), Holmes (1975), and Patton and TAILLEUR (1977). Herald and Wrangel arches and the Lisburne Hills (Fig. 18-2) comprise a fault and fold belt which is younger than the Brooks Range/De Long Mountains; these younger structures are superimposed on and truncate the older ones (Grantz et al. 1975). According to the model, this zone represents a juncture between fold systems of differing ages, the major transport having taken place along the Herald Fault Zone.

The sequence of steps by which this model accounts for the present-day structural relationships is as follows (Fig. 18-3 through 18-6):

(1) In Late Cretaceous time, northward travelling overthrusts culminating in the Brooks Range and De Long Mountains resulted in aggregate longitudinal shortening of at least 130 km. Gravity sliding from a southerly uplifted area (Martin 1970), or backthrusting of the Colville Geosyncline beneath the geanticline (Tailleur 1973) are possible causative mechanisms. Sachs and Strelkov (1961) show that the Pacific and Arctic basins were connected by a seaway across eastern Seward Peninsula at this time (Fig. 18-3).

Figure 18-3. Late Cretaceous tectonic framework of the Bering Strait region.

(2) Easterly directed compressional deformation beginning in latest Cretaceous time produced eastward and northeastward overthrusts along an arcuate front extending from eastern Seward Peninsula to Wrangel Island (Martin 1970, Sainsbury 1972, Patton 1973). Minimum aggregate shortening in the Lisburne Hills uplift was 18 km (Martin 1970). During this phase, the transcurrent faults developed along the eastern and western ends of the De Long Mountains (Martin 1970) in response to the eastward compression, causing the De Long Moun-

tains to assume an arcuate northward bulge (Fig. 18-4). Closure of the seaway between Seward Peninsula and mainland Alaska occurred along the thrust belt (Sachs and Strelkov 1961) at this time.

Figure 18-4. Latest Cretaceous tectonic framework of the Bering Strait region.

(3) In late Cretaceous and early Tertiary time (Fig. 18-5) thrusting ceased, and a trans-current faulting phase began. Strike-slip movement occurred on the Kobuk and Kaltag faults, possibly localized along pre-existing faults or zones of weakness, and the wedge-like region between these faults (Seward Peninsula and Yukon-Koyukuk basin) moved relatively eastward into north-central Alaska. The thrust front extending from Seward Peninsula to Wrangel Island was offset along the offshore extension of the Kobuk fault (Fig. 18-5). The oroclinal pair in the Richardson Mountains far to the east may have been formed at this time as a result of

Figure 18-5. Late Cretaceous/early Tertiary tectonic framework of the Bering Strait region.

right-lateral movement along the porcupine lineament and Tintina fault (Tailleur 1973). Total displacement along the Kaltag fault is approximately 130 km (Patton and Hoare 1968)—enough to account easily for the projected left-lateral offset of the thrust belt north of Seward Peninsula, if we assume a compensating right-lateral offset of similar magnitude along the Kobuk fault.

(4) Subsidence of Norton and Hope basins (Fig. 18-6) probably began in earliest Tertiary time (or

Figure 18-6. Early-middle Tertiary tectonic framework of the Bering Strait region.

shortly before), during the late stages of the transcurrent faulting phase. The subsidence could have begun after the cessation of the major phase of thrusting evidenced in the Lisburne Hills or Herald Arch; the basins could also be a negative tectonic element resulting directly from the compression. Early Tertiary thrust faulting along the western tip of Seward Peninsula (Sainsbury 1972) created a fractured zone along which the new seaway, Bering Strait, would be established in late Miocene time.

(5) Vertical and horizontal adjustments have recurrently affected the region since early Tertiary time. Hope and Norton basins continued to subside, finally accumulating from 3 to 5 km of Paleogene and Neogene sediments in their deepest parts (Grantz et al. 1975, Holmes and Fisher 1979), and some minor Paleogene thrusting took place in eastern Seward Peninsula (Sainsbury 1972). Mild Quaternary uplift occurred in areas of northern Alaska and northeastern Siberia (Stovas 1965). These vertical movements around the margins of Hope basin probably contributed to the formation of some of the steep normal faults which cut the Tertiary basin fill. The presence of the angular unconformity on top of the Neogene (?) unit over the

crest of Kotzebue anticline in southeastern Hope basin indicates that this ridge also experienced relative uplift and erosion in Quaternary time, and some right-lateral movement continued along the Kaltag fault during the Holocene (Grantz 1966, Patton and Hoare 1968).

This discussion has interesting implications for theories of Arctic Basin evolution which require a large suture or fault to extend southward from the Canada basin continental margin across the Chukchi and Bering Sea shelves. Freeland and Dietz (1973) have proposed such a model, stating that evidence for such a fault is provided by the prominent north-south belt of conspicuous magnetic anomalies (Bassinger 1968) recorded in eastern Herald basin.

But Grantz et al. (1975) have traced the Albian and younger Mesozoic beds of the Colville Geosyncline and Herald basin and the Late Cretaceous to early Tertiary deformational zone of folds and faults of the eastern and western structural provinces across this magnetic anomaly belt—evidence which does not support the hypothesized fault zone. The magnetic anomalies first mapped by Bassinger (1968) evidently are related to pre-Cretaceous intrusive bodies similar to those found at many other locations in Alaska and Siberia.

The hypothesis of Freeland and Dietz (1973) would also require that slip along the Kobuk fault be right-lateral. Although evidence to the contrary is convincing (Lathram 1972, Baker 1974), the possibility exists that this broad and complex feature has experienced two episodes of movement in opposite directions.

This is only one (Freeland and Dietz 1973) of a number of hypothetical reconstructions proposed in papers about postulated relative motions of American, Pacific, and Eurasian plates which have been written since the general acceptance of the paradigm of "new global tectonics" (Isacks et al. 1968). Not all the authors of these models seem to have recognized the fact that sea-floor spreading along mid-ocean ridges has global rather than local and regional effects. The various Atlantic, Arctic, and Pacific spreading models produce an interesting, if somewhat confusing, array of possible crustal movements in the area of northern Alaska and northeastern Siberia because of its position with respect to the proposed major plate boundaries (Fig. 18-7). Most of these proposed tectonic movement models are mutually exclusive; many others are completely conjectural and were proposed only in order to solve a problem in some other remote ocean basin.

Figure 18-7. Compilation of proposed crustal movements affecting Alaska and northeastern Siberia (after Lathram 1973).

SUMMARY

The long-term similarity in the geologic histories of northern Alaska and northeastern Siberia and the close relationship between adjacent elements precludes the possibility that any large relative movements have occurred, especially in a north-south direction, between these two land masses. Within the restrictions imposed by this observation and the established chronologies and character of the major tectonic features of this region, a tectonic model can be formulated which accounts for the observed displacements and deformational styles.

Two important assumptions have been corroborated by field observations: the Seward Peninsula is more closely related to the Chukchi Peninsula than to adjacent parts of Alaska, and the Kobuk Trench represents a major transcurrent fault along which the most recent sense of movement has been left-lateral. The model requires eastward movement of the Siber-

ian block toward Alaska, and involves a net transport of approximately 100-150 km.

After the Late Cretaceous development of northward-directed thrust and slide sheets in the De Long Mountains and Brooks Range, strong eastward-directed compressive forces resulted in the formation of a long arcuate thrust belt extending from eastern Seward Peninsula across the Lisburne Hills and Lisburne-Wrangell Arch to Wrangell Island. A long-established seaway connecting the Arctic and Pacific basins across eastern Seward Peninsula was closed by this thrusting and uplift, which involved a minimum eastward displacement of approximately 20 km. Transcurrent faulting occurred at the eastern and western ends of the De Long Mountains in response to this compression.

Continued compression in Late Cretaceous to early Tertiary time resulted in large offsets of approximately 130 km along the Kaltag (right-lateral) and Kobuk (left-lateral) faults, and northeastward bowing and

thrusting of the Lisburne-Wrangel Arch. Subsidence of Hope basin and Norton basin was initiated in the Late Cretaceous or earliest Tertiary. The eastward and northeastward movements in the Lisburne Hills and along the Lisburne-Wrangel Arch produced along the front of the thrust zone a belt of folds and faults (western structural province), which truncated similar features established as a result of earlier northward movements in the De Long Mountains and Brooks Range.

Subsequent horizontal and vertical movements since the early Tertiary have resulted in continued subsidence of Hope basin in the southern Chukchi Sea and Norton basin in the northern Bering Sea, with minor transcurrent motion along the Kobuk and Kaltag faults.

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Section IV

Chemical Oceanography

Donald W. Hood, editor

Some Geochemical Characteristics of Bering Sea Sediments

19

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ABSTRACT

Biogeochemical data are presented for surficial Bering Sea sediments; most are for separate single collections on the southeastern shelf and Norton Sound. Sand-sized sediment predominates, gravel occurs in certain nearshore areas, and mud-sized material is usually a minor component except in Norton Sound adjacent to the Yukon River discharge. In general, the distribution of size fractionation conforms to the present physical environment as this is currently understood and relict and palimpsest sediment is of minor distribution. Southeastern shelf infauna demonstrates a reciprocal relationship: individual organisms are at a maximum in fine sand-sized sediment, but (wet weight) biomass increases in sediment finer and coarser than this. Illite and a glycol-expandable component are the dominant shelf clay minerals ($<2\ \mu\text{m}$), together with chlorite and minor kaolin. Heavy metal contents—especially defined extractable fractions—correlate with fineness of mean grain size; hence contents are, in general, relatively reduced over these shelf areas. Increases in near-bottom particulate contents may be attributed to sediment resuspension.

INTRODUCTION

The work reported in this chapter relates primarily to samples collected on OCS-sponsored cruises: to the southern Bering Sea/Bristol Bay region in 1975, and to Norton Sound in September 1976. These were chemical survey cruises, and, of necessity, the sample grid spacing employed was far too coarse to permit detailed geochemical characterization of any one region. Furthermore, our primary long-term goal has been to contribute to an understanding of the dynamics both of the shelf sediments and of chemicals across the benthic interface. In this respect, not only are the present descriptive data embarrassingly sparse, but interpretation is limited by

insufficient extant information about the physical regime. We can report only inferences about trends and variations. Nevertheless, the sample sets have been collected from little-studied regions of the Alaskan shelf, and the data should form a useful framework for more detailed investigations in the future.

PREVIOUS WORK

The physical oceanographic regime

The general circulation of the Bering Sea was initially described by Ratmanoff (1937), who reported that most of the water flowing in from the north Pacific entered through passes in the central and western Aleutian Islands and exited northwards by way of the Bering Strait. Later workers have emphasized that the induced Pacific circulation is restricted to the region adjacent to the Aleutian chain. Thus, from observations in the vicinity of Unimak Pass, Barnes and Thompson (1938) described a current structure northward from the passes basically defined by the shelf topography. It may be inferred from this and other early work that the north-northwest flow over the outer continental shelf tends to isolate the southeastern shelf region from the surface waters of the southern and central Bering Sea.

Bristol Bay has received more attention than any other region of the Bering Sea. Dodimead et al. (1963) summarized circulation investigations

made over the previous 30 years or so. Hebard (1959) and Takenouti and Ohtani (1974) have described a cyclonic circulation on the southeastern shelf supposedly in response to regional meteorological, tidal, and topographic control. Intensive recent investigations have revised these concepts somewhat. Through the ice-free season Bristol Bay waters can be basically classified into three water types: coastal water lies shoreward of an oceanographic front located near the 50-m isobath; shelf water occupies the region approximately between the 50-m and 100-m isobaths; Alaskan Stream/Bering Sea water intrudes into the southeastern basin region (Coachman and Charnell 1979).

Kinder (1977) and Schumacher et al. (1979) have described the structural front which, coinciding approximately with the 40-50-m depth contours, separates the well-mixed coastal and two-layered central shelf domains. This feature is characterized by marked vertical property gradient changes, has a typical width of 5-10 km, and is generally located where the input of buoyant energy balances tidally generated mixing energy; the front occurs where the height of the bottom mixed layer closely approximates the water depth so that tidal mixing occurs throughout the (almost vertically homogeneous) column.

Recent research (especially by the PROBES group) has provided a basic description of the outer shelf region (between Unimak Island and the Pribilofs) between the shelf break—around 150 m—and approximately the 100-m depth contour. This region, bounded by two oceanographic fronts, appears to constitute a transition zone between the Bering Sea source and deep shelf water. The inner boundary front (80-100 m depth, around 100-130 km from the shelf break) separates the two-layer shelf region from the transition zone which is basically three-layered: upper and bottom zones subject to wind and tidally generated mixing respectively, separated by a region of relatively low energy.

This front extends from surface to bottom and appears to migrate into deeper water in the winter, when wind mixing operates to greater depths (50-60 m). The outer front is located mainly over the shelf break and forms a partial barrier between oceanic source and shelf waters. Since this structure does not extend to the bottom, Bering Sea water flows landward along the bottom of the outer shelf. Schumacher et al. (1979), Coachman and Charnell (1979), and Coachman (1979) report no appreciable net circulation in the central shelf region; but tidal mixing operates to 40-50 m above the bottom and currents of the order of 50 cm/sec have been ob-

served. The other effective mixing mode—wind energy—operates to 10-20 m in spring and summer and, as noted, considerably deeper in winter.

Surficial sediment studies

Creager and McManus (1967) briefly described the bottom sediments of the northern Bering Sea and summarized previous cruises and work to around 1965. McManus et al. (1974) thoroughly characterized a central-northern region. There have been many limited regional studies, particularly in the southeastern Bristol Bay region (see summary of Askren 1972) and the Gulf of Anadyr (considerable Soviet literature). Sharma et al. (1974) have considered the use of satellite imagery to delineate sediment transport within Bristol Bay. A comprehensive bibliography has recently been compiled by Naidu (1977).

PROCEDURES

Surficial sediment samples from the southern Bering Sea for which size fractionation, clay mineralogy, heavy metal, and infauna data are given in this report were mostly collected in June 1975, using a Haps (Kannevorff and Nicolaisen 1973) stainless steel corer and van Veen grab. Additional samples for the size analysis program were collected on subsequent NOAA/OCS-sponsored cruises in the same area. Suspended sediment samples were obtained on the former cruise by in-line filtering water from rosette-mounted drop-top Niskin bottles through a Nuclepore membrane. Since this would have been an unsuitable procedure for obtaining column samples to determine soluble trace constituents, a separate sampling program for these was conducted aboard the U.S.S.R. hydromet vessel *Volna* in July-August 1977. The stations occupied on this joint U.S.-U.S.S.R. expedition did not, unfortunately, coincide with stations of the earlier cruises. Bottom sediment samples were collected in Norton Sound (September, 1976) using the same techniques as for the southeastern Bering/Bristol Bay region. Sources of the additional clay mineralogy data for the northern Bering Sea have been given by Naidu et al. 1977.

Granulometric analysis has been by conventional sieving-pipetting procedures (Folk 1968) with generation of Inman (1952) statistical parameters through a modified computer routine after Creager et al. (1962). For the southeastern Bering Sea shelf samples the phi-grade scale of Krumbein (1936) and the Wentworth (1922) size limits and class terms have been used; also the size terms of Folk (1954), based on a classification scheme that emphasizes gravel,

since the proportion of gravel is in part a function of the highest current velocity at the time of deposition. Only major categories (gravel, sand, mud) were determined for the surficial sediment samples collected from Norton Sound. Clay mineral analysis essentially followed the procedures given by Naidu et al. (1971) and Mowatt et al. (1974). All clay mineral quantifications in this study are based on the method of Biscaye (1965) and are, at best, semi-quantitative.

Total analysis for elemental constituents of deposited and particulate sediment has been performed by D. E. Robertson using neutron activation: Mn, V, and Al by "rabbit" irradiation, and the other elements using techniques given by Robertson and Carpenter (1974). Chemical extracts of sediments were obtained using 25 percent v/v acetic acid as mandated by the contracting agency. Metal concentrations thus determined are lower than would have been given by the more conventional mixed acid-reducing treatment; but the latter has only the virtue of more universal use, since several studies (e.g., Luoma and Jenne 1976) have shown little actual correlation between "bioavailability," as evidenced by measured uptake, and various chemical leaching treatments. Soluble copper and lead contents of seawater were obtained using anodic stripping voltammetry (Heggie and Burrell 1976).

Identification and quantification of the southern Bering Sea infauna have been performed by the University of Alaska Marine Sorting Center.

SEDIMENT DISTRIBUTION

The distributions of mean phi, sorting (standard deviation), grain-size mode for sand, and the percent abundance of gravel, sand, silt, and clay across the southeastern Bering Sea continental shelf are shown in Figs. 19-1 to 19-7. Such maps of distributions of (surficial) sediment characteristics can be used to indicate source areas, processes of transportation and deposition, and environments of deposition (K. Tommos, unpublished data); only a brief summary is given here.

Sand-sized sediments, in the classification scheme of Folk (1954), dominate this region of the Bering Sea shelf; in Wentworth class terms, coarse silt is also regionally significant. The relative abundance of sand ranges from about 20 percent to almost 100 percent (average content, 68 percent). Samples collected along the Alaska mainland, at an average depth of about 30 m, consist of >90 percent sand, and an 80-percent sand isopleth generally follows the 45-m isobath. The silt content in this region is low—less than 10 percent. It was found that isopleths of

Figure 19-1. Southeastern Bering Sea shelf: Grain-size mean (ϕ).

Figure 19-2. Southeastern Bering Sea shelf: Sediment sorting.

sediment textural parameters tend to follow bathymetric contours on the inner shelf: sediment finer than 125 μm (3 ϕ) does not occur in depths of <35 m and generally the mean size decreases uniformly below this depth. Sediments coarser than 250 μm (2 ϕ) are restricted to depths of <50 m, suggesting that resuspension and transport of fine-grain sediments occur in this zone—conclusions which support the findings of Askren (1972). This sedimentological pattern is compatible with recent concepts of the local physical oceanographic regime. The high

Figure 19-3. Southeastern Bering Sea shelf: Grain-size mode (μm) for the sand fraction.

Figure 19-4. Southeastern Bering Sea shelf: Abundance of gravel in surface sediment (%).

Figure 19-5. Southeastern Bering Sea shelf: Abundance of sand-sized sediment (%).

Figure 19-6. Southeastern Bering Sea shelf: Abundance of silt-sized sediment (%).

sand content, well-sorted character, and consistent increase in mean size in depths shallower than approximately 50 m reflect the proximity to the mainland sediment sources and the subsequent influence of wind waves and tidal, storm, and coastal currents. The sandy gravel, gravelly sand, and sand found in this southeastern shelf region are apparently being supplied by the bordering rivers: the Egegik, Naknek, Kvichak, Nushagak, and Kuskokwim Rivers are major contributors. Poorly to very poorly sorted sediment is present only close to shore.

Gravel-sized material was obtained mostly in nearshore areas, especially in sediment from Bristol Bay proper, Kuskokwim Bay, and Unimak Pass. These sediments probably represent reworked fluvial or beach deposits, or, in isolated cases, relict or palimpsest deposits. The relative amounts of the silt component in the sediments over the entire southeastern Bering shelf region range from 0.02 to 50 percent with a mean content of around 20 percent. This component is dominant in the low-energy environment region of this southern shelf; the largest

Figure 19-7. Southeastern Bering Sea shelf: Abundance of clay-sized sediment (%).

concentrations, however, occur in a number of isolated patches (K. Tommos, unpublished data). The mean concentration of clay-sized sediment is around 7 percent, with greater (but never dominant) contents north of Unimak Pass and St. Matthew Island, adjacent to St. George Island, and generally in the northwest part of this shelf region.

Large gradients in grain size, sorting, and sand and silt content which are evident adjacent to the 40-50 m zone are believed to be the result of, or at least influenced by, the presence of the seasonally stable oceanographic front which here separates the coastal and central shelf domains. The pronounced gradients in turbulence and hence transport efficiency result in deposition of coarser-grained sediment landward and finer-grained sediment seaward of this zone. Across the remainder of the shelf, the relative amount of the sand component decreases discontinuously. Sediments having a larger mean ϕ (very fine sand) and improved sorting were collected near the shelf edge at around 150 m. This supports the presence of the permanent current proposed by Barnes and Thompson (1938). More recent investigations (Coachman and Charnell 1979) have emphasized the presence of a very low velocity northwesterly drift along the outer shelf. But Coachman (1979) has noted that sporadic eddies along the shelf edge may produce local transient flow of basin water onto the shelf. The effect of this on the energy regime could account for the increasing mean grain size and improved sorting in this zone.

At this time it is possible to give only a general description of the textural variations in the surface

sediments of the southeastern Bering Sea shelf. The preliminary assumption can be made that this sediment is in dynamic equilibrium with the present energy environment. Emery (1968) considered that relict sediment covers most of the shallower portions of the Bering Sea, but this has been generally disputed by more recent workers (e.g., Askren 1972, Knebel 1972, Sharma 1972). There are as yet no satisfactory explanations for those restricted regions still defined as relict or palimpsest. These are frequently interpreted as fluvial material deposited during periods of lowered sea level. McManus et al. (1974) summarized contemporary studies indicating that the Yukon River previously flowed south of St. Lawrence Bank, so that fine sediment from this source would have been deposited in the present southern Bering Sea. The areal plots of sediment texture given here generally show a decrease in grain size seaward across the shelf agreeing with the graded shelf model of Swift (1970). The sediment finer than 250 μm (2ϕ) and the deterioration of sorting across the shelf reflect the decrease in the energy of dispersal mechanisms with water depth and distance from shore. There are, however, significant interruptions in this trend, suggesting that the surficial sediments of the southern Bering Sea continental shelf are still approaching equilibrium with the present physical environment.

Only primary size fractionation data (mud, sand, and gravel categories) are presently available for the Norton Sound surface sediment samples as shown in Figs. 19-8 to 19-10. Sand-sized material predomi-

Figure 19-8. Norton Sound: Abundance of mud-sized sediment (%).

Figure 19-9. Norton Sound: Abundance of sand-sized sediment (%).

Figure 19-10. Norton Sound: Weight (%) abundance of gravel in surficial sediment.

nates in the outer reaches of the sound, as on the southeastern shelf discussed above. Sand is also the major component of the surficial sediment within Norton Bay. Within the embayment adjacent to Unalakleet, however, and over a considerable portion of the central region of Norton Sound, mud (clay-plus silt-sized sediment) constitutes the dominant category. The latter deposit is clearly attributable to Yukon River discharge, although most of the immense quantity of sediment derived from this source is transported northwards through the Bering Strait. Gravel is dominant adjacent to the shoreline off Port Clarence.

INTERACTIONS BETWEEN BENTHOS AND SEDIMENT SUBSTRATE

Scatter diagrams for total number of individuals and wet weight (m^{-2}) for each of the 25 southern Bering Sea stations having paired benthos-sediment size fractionation values are given as Figs. 19-11 and 19-12. These data demonstrate a reciprocal relationship with the largest total number of individuals (on a unit area basis) and the lowest total wet weight occurring at localities having grain-size modes between 125 and 149 μm (fine sand); larger weights of macrobenthos tended to occur where either coarser or finer sediment predominated. Since sediment particles between 100 and 200 μm are among the first to move under current stress (Sternberg 1972) and

tend to be well sorted as a consequence, it would appear likely that the large number of small individuals at fine-sand stations is the result of tidal current action.

Correlation coefficients between numbers and weights (m^{-2}) of each species of sorted macrobenthos (where present at >2.5 percent) have been determined (Hoskin 1978). Of 138 species, 16 have a correlation coefficient of 0.5 or greater and these may be conveniently considered by feeding mode. Scavengers (e.g., the amphipod *Paraphoxus* sp., Fig. 19-13) increase in abundance with increasing coarseness of grain-size, and there appears to be a threshold in substrate grain size since neither this organism nor *Echinarachnius parma* were found where grain-size modes were $<149 \mu\text{m}$.

For three predators (polychaetes, *Nephtys caeca*, and *N. longasetosa*) both numbers and weights increased with increasing coarseness of grain-size modes in their substrate. This trend was reversed for the gastropod predator *Cylichna alba*.

The deposit-feeding polychaetes *Ophelia limacina* and *Spio filicornis* (Fig. 19-14) were also observed to increase in number and biomass with increasing grain size, and there may be a particle-size threshold for the former since it was not found in this sampling in substrates having grain-size modes $<105 \mu\text{m}$. For other deposit-feeding polychaetes identified, either the data were inconclusive (*Ampharete arctica* and *Travisia forbesii*), or no relationship between abun-

Figure 19-11. Scatter plot for total number of macrobenthos individuals (m^{-2}) vs. grain-size mode for southeastern Bering Sea shelf.

Figure 19-12. Scatter plot for total wet weight of macrobenthos (m^{-2}) vs. grain-size modes for southeastern Bering Sea shelf.

dance and substrate modal grain size was apparent (*Myrioichete heeri* and the Terebellids). The filter-feeding amphipod *Haustorius eous* was not found in substrates with grain-size modes finer than 149 μm . The abundance of this organism decreased with increasing sediment coarseness, but the relationship between weight per unit area and substrate grain-size modes is unclear.

It seems clear that characterization of the substrate by primary (dispersed) grain-size analysis can provide at best only an imperfect description of habitat.

Johnson (1974) has well emphasized the importance of mixed-phase aggregates to benthos distributions—parameters which cannot be determined using conventional size fractionation techniques. Nevertheless the data summarized here (Hoskin 1978) provide a partial first-order description of animal-sediment relationships in the southern Bering Sea.

CLAY MINERALOGY

The clay minerals of the southeastern Bering Sea shelf consist predominantly of an expandable component (40-70 percent: Fig. 19-15) and illite (20-50 percent: Fig. 19-16) together with chlorite (30-40 percent). Kaolinite is present generally in amounts less than 10 percent. There are also indications of amorphous material in concentrations greater than have been found elsewhere on the Alaskan shelves.

Figure 19-13. Scatter plot for number of individuals of the scavenger amphipod *Paraphoxus* sp. (m^{-2}) against grain-size mode for southeastern Bering Sea shelf.

Figure 19-14. Scatter plot for wet weight of the deposit feeding polychaete *Spio filicornis* (m^{-2}) against grain-size mode for southeastern Bering Sea shelf.

Figure 19-15. Distribution of expandable mineral component (%) in $<2 \mu\text{m}$ size fraction of southeastern Bering Sea sediments.

The results of potassium and magnesium saturation of the (glycol) expandable mineral component suggest the latter to be degraded illite. The $<1 \mu\text{m}$ (e.s.d.) fraction shows less illite but an increased expandable component as compared with the $<2 \mu\text{m}$ fraction (Naidu and Mowatt 1977). Expandable material may represent contributions from bordering volcanic regions; the kaolinite component presumably is introduced by way of the Kuskokwim-Nushagak and other river systems.

Fig. 19-17 illustrates the primary distribution of clay minerals in the north Bering Sea region, re-

presenting pooled information from a number of cruises and sources (A. S. Naidu, unpublished data). Naidu et al. (1977) have also recently published the results of a survey of some 80 surficial samples from this area. Again illite is invariably the predominant clay mineral present in the $<2 \mu\text{m}$ e.s.d. size class. Similarly, subordinate amounts of a glycol-expandable mineral (10-40 percent), chlorite (18-32 percent), and kaolinite (6-18 percent) are present. Significant amounts of the expandable component appear to be degraded illite, and enhanced amounts (30-40 percent) were observed south of St. Lawrence

Figure 19-16. Distribution of illite (%) in $<2 \mu\text{m}$ size fraction of surficial sediments of southeastern Bering Sea.

Island. The region extending off the Yukon River north up to the eastern margin of the Chirikov Basin, the western portion of Norton Sound, and the Shpanberg Strait demonstrate significantly higher kaolinite/chlorite ratios. This is attributable to a local terrigenous source and current dispersal. The presence of notably high concentrations of expandable component south of St. Lawrence Island is difficult to explain. Possibly clays of the region are relict and have been formed by alteration of local volcanic debris. The distribution patterns of clay minerals, together with corresponding information

from north of the Bering Strait not considered above, support the common thesis that the Chukchi Sea is the major depositional site for clay-size sediment from the Yukon River.

SEDIMENT HEAVY METALS

Since the specialized sampler required to obtain contamination-free surficial sediment samples could not operate in sandy or coarser sediment, only a limited set of samples was available to us for this phase of the study. Both total and extractable (i.e.,

Figure 19-17. Distribution of illite and expandable clay mineral component and kaolinite/chlorite ratio of $<2 \mu\text{m}$ size fraction of northern Bering Sea region.

“available”) heavy metal contents were determined for the south Bering Sea and Bristol Bay. Whole-rock sample concentrations of a wide range of trace metals have been given by Robertson (1977). Mean extractable and total contents, and correlations of the latter, are given in Table 19-1. Fig. 19-18 illustrates the areal distribution of total vanadium ($\mu\text{g/g}$) as an example.

These heavy metal contents are low—much lower than mean values reported for polluted coastal regions, but also lower than has been determined for surficial Gulf of Alaska sediments (as part of the same survey program: Burrell 1978). Robertson (1977) has drawn attention to the negative cor-

relation with calcium (Table 19-1): calcareous sediments demonstrate very low heavy-metal contents. But it is suggested here that the generally coarser character of the sediment as compared with, for example, the eastern Gulf of Alaska, is also a major factor: finer-grained sediments generally contain higher concentrations of sorbed metals. This is less readily apparent from these data because the metals for which both extractable and total data exist—manganese and iron—occur as primary oxide precipitates and coatings. Figs. 19-19 and 19-20 show distributions of iron in this region: extractable contents (25 percent v/v acetic acid treatment) are some order of magnitude less than totals. The coefficients given in Table 19-1 primarily appear to demonstrate highly significant (0.1) correlations for structural alumina-silicate elements: Al, Fe, Ca, Co.

Although lack of suitable facilities prevented comprehensive collection of water-column samples for trace elemental analysis, a limited number of particulate sediment samples were collected concurrently with the bottom sediment samples described above and these have been analyzed by D. E. Robertson (various unpublished reports). Mean values ($\mu\text{g/l}$) for surface and near-bottom samples are given in Table 19-2. Robertson (1977) has suggested that the particulate values for manganese and aluminum are relatively high and for vanadium relatively low for an open-ocean environment.

Near-bottom total Al, Fe, Mn, and V values are also significantly higher than those obtained for surface samples. This characteristic should perhaps be considered in conjunction with the mean soluble data for copper and lead in central Bering Sea waters

Figure 19-18. Distribution of total vanadium (ppm) in surficial sediments of southeastern Bering Sea shelf.

TABLE 19-1

Mean extractable and total metal concentrations and correlation coefficients of total contents for surficial sediments of the southern Bering Sea

Contents			Correlation Coefficients							
Extractable	Total		Al	Ca	Fe	Mn	V	Cr	Co	As
0.14 ± .05%	6.09 ± .96 %	Al								
	2.86 ± .9 %	Ca	.847							
	2.84 ± .55 %	Fe	.915	.782						
	494 ± 199 ppm	Mn	.733	.361	.715					
	93 ± 17 ppm	V	.880	.569	.811	.948				
	64 ± 21 ppm	Cr	.570	.774	.648	.337	.435			
	10 ± 2 ppm	Co	.863	.738	.950	.629	.745	.596		
	3.6 ± 1.5 ppm	As	.466	.338	.713	.642	.582	.534	.707	
	0.6 ± .15 ppm	Sb	.500	.544	.658	.495	.474	.679	.706	.661
< 2.5 ppm	Ni									
11 ± 6 ppm	Zn									

also given in Table 19-2 (these latter data are considered more fully by Burrell, Chapter 21, this volume; Burrell 1978; and especially Heggie, 1980, and unpublished data). Near-bottom enhancement of particulate Al (in w/v units) would suggest bottom sediment resuspension, and such a mechanism could account for apparent local increases in the concentrations of Fe, Mn, and V at the base of the water column (Robertson, unpublished report). Such could also be extended to include enhancements in "soluble" contents—for example, the copper and lead of Table 19-2; i.e., enhancements due to the presence of particulate material less than the nominal 0.4 μ m filter pore size (Burrell 1978). Fig. 19-21 illustrates the distribution of surface soluble (passing 0.4 μ m membrane) copper and gives data for near-bottom samples from within the 100-m shelf contour. The mean bottom-water copper concentration given in Table 19-2 is for this shelf suite. It has been argued (Heggie, 1980) that the shallow-water copper gradient (negative away from the benthic boundary into the column) may be attributable to a flux of, in this case, remobilized copper from the sediments. We have previously demonstrated such a transport under estuarine conditions for copper (Heggie and Burrell 1980) and manganese (Owens et al. 1979), and such a process over a highly productive shelf region is certainly feasible. Data for the distribution of

particulate manganese over the shelf are unfortunately too sparse either to support or refute this argument, and this important topic awaits specific investigation.

Figure 19-19. Content (%) of extractable iron from southeastern Bering Sea sediment.

Figure 19-20. Total iron content (%) of southeastern Bering Sea sediment.

The distribution of size-fractionated sediment and the mineralogy of the clay-sized component within Norton Sound has been briefly noted above. The character of the surficial sediment in this area is influenced to a considerable degree by the Yukon

Figure 19-21. Soluble (<0.4 μm) copper contents of surface waters of central Bering Sea (D. T. Heggie, unpublished data). The second value given for shelf samples shows corresponding near-bottom values.

TABLE 19-2

Some mean elemental particulate and soluble contents of Bering Sea water
(from Robertson, Heggie, unpublished data)

Element		n	Surface	n	Bottom
A. Particulate sediment					
Al	$\mu\text{g/l}$	7	$17.9 \pm .2$	17	$39.0 \pm .3$
Fe	$\mu\text{g/l}$	4	$4.13 \pm .17$	10	$15.8 \pm .5$
Mn	$\mu\text{g/l}$	7	$0.33 \pm .03$	17	$0.83 \pm .07$
V	$\mu\text{g/l}$	-	<16	16	92 ± 19
As	$\mu\text{g/l}$	4	12.3 ± 1.8	10	11.9 ± 1.7
Co	$\mu\text{g/l}$	4	$3.7 \pm .09$	10	$5.4 \pm .08$
Sb	$\mu\text{g/l}$	4	$1.03 \pm .21$	10	$0.88 \pm .27$
B. Soluble concentrations					
Cu	$\mu\text{g/l}$	23	$0.41 \pm .22$	13	$0.69 \pm .38$
Pb	$\mu\text{g/l}$	23	$0.20 \pm .04$	13	$0.35 \pm .10$

River discharge. Table 19-3 demonstrates the close correlation of extractable contents of heavy metals from the sediment with the weight fraction of fine-grained sediment—i.e., with available solid-phase surface area.

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TABLE 19-3

Mean extraction heavy metal concentrations and correlation coefficients for Norton Sound surficial sediments

Metal			Correlation coefficients			
			% Mud	Fe	Zn	Ni
0.2	%	Fe	0.86			
0.01	%	Mn	0.53	0.27		
5.3	ppm	Zn	0.67	0.71		
2.2	ppm	Ni	0.73	0.98	0.76	
0.7	ppm	Cu	0.62	0.84	0.84	0.85
< 0.1	ppm	Cd

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The Distribution and Elemental Composition of Suspended Particulate Matter in Norton Sound and the Northeastern Bering Sea Shelf: Implications for Mn and Zn Recycling in Coastal Waters

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ABSTRACT

The distribution and elemental composition of suspended particulate matter in Norton Sound and the northeastern Bering Sea shelf were studied in July 1979. Samples were analyzed for total suspended matter and particulate C, N, Mg, Al, Si, K, Ca, Ti, Cr, Mn, Fe, Ni, Cu, and Zn. The results show that the bulk of suspended material in Norton Sound consists of sedimentary material discharged from the Yukon River and resuspended bottom sediments. The Yukon River material enters the sound from the southwest, is transported north and northeast around the perimeter of the sound, and exits from the northwest.

The concentrations of the major and trace elements in the particulate matter and their elemental ratios with aluminum indicate that: K, Ca, Ti, Cr, Fe, Ni, and Cu are primarily associated with aluminosilicate material derived from the Yukon River and resuspended sediments, and C and N are primarily associated with terrestrial organic material in estuarine samples and marine organic material in offshore samples. Significant enrichments of Mn and Zn, observed in the offshore samples, are attributed to Mn recycling in the sediments followed by precipitation of Mn onto particulate phases in the water column, with the Mn oxyhydroxides scavenging Zn.

INTRODUCTION

Particles suspended in seawater play a major role in regulating the chemical forms, distributions, and deposition of many of its constituents. This is particularly true in coastal waters where dissolved and

particulate matter in runoff from rivers interacts with seawater. Particles in coastal waters are the result of continuing physical, chemical, biological, and geological processes and are the precursors of marine sediments. These processes may include the supply of inorganic and organic substances from river runoff, aeolian fallout and coastal erosion, resuspension of previously deposited sediments, biological production of organic materials, and chemical adsorption-desorption and flocculation processes. Variations in the composition of particles in suspension can be sensitive indicators of such processes. In this chapter we present the results of a survey of the distribution and elemental composition of suspended material in Norton Sound and the northeastern Bering Sea shelf. The results are related to known patterns of water circulation and previously published information on the chemical composition of suspended material from the Yukon River. We present arguments to support the position that Mn and possibly Zn undergo dissolved-to-particulate phase changes as a result of interactions between suspended matter and dissolved species within Norton Sound.

HISTORICAL BACKGROUND

Previous work on suspended matter in Norton Sound has been limited to studies of LANDSAT photographs and distributions of suspended matter. Sharma et al. (1974) used density-sliced LANDSAT photographs and sea-truth measurements to study distributions of suspended matter in Norton Sound during the late summer of 1973. Concentrations of suspended matter were highest near the mouth of the Yukon River (range: 2-8 mg/l) and in Norton Bay (range: 3-4 mg/l), in the northeast corner of the sound. The authors postulated that the general pattern of cyclonic circulation in the sound caused suspended material to be transported to the north and northeast along the coast. The authors also noted that unusually high concentrations of particulate matter (>9.0 mg/l) were observed throughout the water column in the region approximately 30 km south-southwest of Nome. They suggested that this plume could have been a detached portion of the Yukon River plume which was isolated by tidal pulsation.

Cacchione and Drake (1979) combined surveys of suspended matter during September and October 1976 and July 1977 with deployments of a tripod (GEOPROBE) containing instruments designed to measure bottom currents, pressure, temperature, and light transmission and scattering to study dispersal patterns of suspended matter in Norton Sound. They described the transport of suspended materials as dominated by distinctly different quiescent and storm regimes. The quiescent regime was characterized by relatively low levels of sediment transport caused by tides and mean flow to the north and northeast, augmented by surface waves during spring tides. The authors stated that in this period, much of the fine-grained suspended matter present over the prodelta was resuspended at shallow depths during spring tide and transported northward with the mean current. The storm regime, which occurs during the months of September through November, was characterized by strong southerly and southwesterly winds which generate waves with heights of 1-3 m and periods of 8-11 sec. The storm events cause near-bottom shear velocities in excess of that required for resuspension of bottom sediments and, as a result, more than 50 percent of the sediment transport occurs during this regime.

Although there is no background information on the chemistry of suspended matter in Norton Sound, extensive studies of trace-metal partitioning in various phases of Yukon River materials were conducted by Gibbs (1973, 1977). He concluded that transition

metals associated with oxyhydroxide coatings and crystalline phases comprised the major fraction (72-91 percent) of riverine transition metal transport to the sea. Particulate organic phases contained the next largest fraction (3-16 percent of the total). Metals in solution and metals sorbed to particulate materials made up the remainder (5-15 percent of the total).

Figure 20-1. Physiographic setting of Norton Sound showing: (a) locations of suspended matter stations, 7-18 July 1979 (b) bathymetry in meters.

THE STUDY REGION

Norton Sound is a shallow embayment of the Bering Sea in the central region of the west coast of Alaska (Fig. 20-1). It extends east-west about 220 km and north-south about 150 km. The Yukon River, which flows into the southwest quadrant of the embayment, is the major source of fresh water and suspended matter to the sound as well as to the entire eastern Bering Sea shelf. Its annual load of suspended matter, 88×10^6 tons, ranks 18th among the major rivers of the world (Inman and Nordstrom 1971). The annual discharge curve for the Yukon River (Fig. 20-2) is unimodal with peak flow occurring during June and low flow conditions persisting throughout the winter months. Additional lesser freshwater influx into the sound occurs along the coastline east of the Yukon River Delta and along the northern coast.

Water circulation in the vicinity of Norton Sound has been described by several authors (Coachman et al. 1975, Muench and Ahlnäs 1976, Muench et al. Chapter 6, this volume). The shelf water west of Norton Sound, the Alaskan coastal water, has a net northward flow of about 1.5×10^6 m³/sec. About one-third of this flow passes between St. Lawrence Island and the mouth of Norton Sound. This flow induces the cyclonic water circulation inside Norton Sound. The intensity of the cyclonic flow appears to be affected by local winds and by freshwater runoff. The eastern half of the sound is characterized by two vertically well-mixed layers. The upper layer contains runoff water from the coastal rivers; the lower layer contains cold, dense residual water formed during the previous winter. Both water masses follow the general pattern of cyclonic flow in the region, although much more sluggishly than surface and bottom waters further to the west.

The distribution of sediments in Norton Sound has been summarized (McManus et al. 1974, Sharma 1974, Nelson and Creager 1977, and McManus et al. 1977). In the central and southern regions, the sediments consist of very fine grained sands and silts which are modern. In the northern region, silty sands predominate everywhere except for a narrow strip along the coast between Cape Nome and Cape Douglas. Here, coarse sands and gravels predominate because bottom currents have caused almost complete erosion of the fine-grained sediments. Approximately one-half to two-thirds of the sediment load of the Yukon River is deposited as a band of sediments extending from the Yukon River Delta northward and eastward around the perimeter of the

sound. The remaining sediment load of the Yukon River is transported to the north through Bering Strait and deposited in the Chukchi Sea.

Figure 20-2. Monthly means and ranges for Yukon River discharge. Data compiled from U.S.G.S. streamflow obtained at Pilot Station (located approximately 200 km upstream from the river mouth) for period of record: 1975-78.

EXPERIMENTAL PROCEDURES

Sampling methods

In order to obtain information about the distribution and composition of suspended matter in Norton Sound, samples were collected as part of an interdisciplinary survey of the region (7-18 July 1979). Fig. 20-1 shows the locations of the stations occupied during the survey. Water samples were collected in General Oceanics Model 1070 10-l PVC Top Drop Niskin bottles from the surface and 5 m above the bottom, and from intermediate depths along two vertical sections spanning the length of the

sound (Transect I: Stations 3, 4, 8, 11, 15, 20, 29, 38, 43, and 48; Transect II: Stations 1, 9, 10, 17, 18, 22, 25, 27, 39, 42, and 49). To avoid loss of rapidly settling particles (Gardner 1977, Calvert and McCartney 1979), aliquots from each Niskin bottle were rapidly withdrawn (within 10 to 15 minutes of collection) and vacuum-filtered through preweighed 0.4- μm pore-size Nuclepore polycarbonate filters (47 mm in diameter for suspended matter concentration determination and 25 mm in diameter for elemental analyses other than C and N) and precombusted 0.45- μm pore-size Sela silver filters (25 mm in diameter for C and N analyses). All samples were rinsed with three 10-ml aliquots of deionized membrane-filtered water (adjusted to pH 8.0), placed in individual polycarbonate petri dishes with lids slightly ajar for a 24-hour desiccation period over sodium hydroxide, and then sealed and stored for subsequent laboratory analysis.

Temperature and salinity data were obtained with a Plessey Model 9040 CTD system equipped with a Model 8400 data logger. This system sampled several times per second for simultaneous values of conductivity, temperature, and depth. The data were averaged to provide 1-m temperature and salinity values, from which sigma-t was computed. The sigma-t values are accurate to ± 0.02 sigma-t units.

Analytical methods

Total concentrations of suspended matter were determined gravimetrically. The weighing precision ($2\sigma = \pm 0.011$ mg) and volume reading error (± 10 ml) yield a combined coefficient of variation in suspended matter concentration of approximately 1 percent. This variability is probably overshadowed, however, by that associated with the sampling precision as reported by Feely et al. (1979) for other Alaskan coastal waters, where the relative standard deviation ranged from 5 to 25 percent.

The major (Mg, Al, Si, K, Ca, Ti, and Fe) and trace (Cr, Mn, Ni, Cu, and Zn) elements in the suspended matter were determined by x-ray secondary-emission (fluorescence) spectrometry using a Kevex Model 0810A-5100 x-ray energy spectrometer and the thin-film technique (Baker and Piper 1976). A silver x-ray tube (operated at 50 kV, 40 mA) was used to excite a sequence of secondary targets (Fe target for Mg through Cr; Se target for the remaining elements) which efficiently fluoresced the range of elements in each sample. Standards were prepared from suspensions of finely ground U.S.G.S. Standard Rocks (W-1, BCR-1, AGV-1, and GSP-1;

90 percent by volume were less than 15 μm in diameter as determined by scanning electron microscopy) collected on Nuclepore filters identical to those used for sample acquisition. At a filter loading of 325 $\mu\text{g}/\text{cm}^2$ the determination limits (three times the minimum detection limits) were less than 0.25 percent and 13 ppm for the major and trace elements, respectively. The relative standard deviations resulting from 10 replicate analyses of a sample with a similar weight distribution were less than 3 percent for major elements and less than 8 percent for trace elements.

The amorphous Mn and Zn in the poorly structured oxyhydroxide phase of selected suspended matter samples were determined by the method of Bolger et al. (1978). Desiccated samples were leached with 5 ml of 25 percent (v/v) Ultrex acetic acid at room temperature for two hours. The resulting supernate was filtered through an acid-cleaned polypropylene-glass apparatus containing a 0.4- μm Nuclepore filter. The residue was rinsed with quartz-distilled water, then filtered; the supernate was combined with the original supernate, acidified with 0.5 ml of concentrated Ultrex HCl, and stored in an acid-cleaned polyethylene bottle. The Mn and Zn in this solution (weak-acid-soluble) were analyzed by flameless atomic absorption procedures using standard addition methods. The remaining solid suspended matter (weak-acid-insoluble) was dissolved in an Ultrex HCl-HNO₃-HF matrix following Eggemann and Betzer (1976) and analyzed for Mn and Zn in a similar manner.

Analysis of total particulate carbon and nitrogen in the suspended matter was performed with a Hewlett Packard Model 185B CHN analyzer. In this procedure, particulate carbon and nitrogen compounds were combusted to CO₂ and N₂ (micro Pregl-Dumas method), chromatographed on Poropak Q, and detected sequentially with a thermal conductivity detector following a modification of the procedure outlined by Sharp (1974).¹ NBS acetanilide was used for standardization. Analyses of replicate surface samples yield relative standard deviations ranging from 2 percent to 10 percent for carbon and 7 percent to 14 percent for nitrogen.

¹ Because the silver filters used for sample acquisition could not be accurately weighed, carbon and nitrogen data were determined on a mass-per-volume basis. Weight percent data for carbon and nitrogen were obtained by comparison with suspended matter loadings obtained with the 47-mm Nuclepore filters.

Figure 20-3. Distribution of: (a) salinity (b) temperature (c) sigma-t (d) total suspended matter at the surface in Norton Sound, 7-18 July 1979.

RESULTS AND INTERPRETATIONS

Distribution and transport of suspended matter

Figs. 20-3 and 20-4 show the distributions of salinity, temperature, sigma-t, and total suspended matter at the surface and 5 m above the bottom for the July 1979 cruise in Norton Sound. As shown in Fig. 20-3, surface distributions of particulate matter were dominated by the discharge of sedimentary ma-

terial from the Yukon River. Surface concentrations of suspended matter were highest near the mouth of the Yukon River, where values ranging between 100 and 154 mg/l were observed. The Yukon River plume (as indicated by the 5.0-mg/l isopleth) extended to the north and northeast across the length of the sound. Another portion of the plume with lesser concentrations of suspended matter (1.0-2.7 mg/l) extended north and northwest to a point about 20 km southwest of Cape Rodney. Both portions

Figure 20-4. Distribution of: (a) salinity (b) temperature (c) sigma-t (d) total suspended matter at 5 m above the bottom in Norton Sound, 7-18 July 1979.

appear to have originated from the Yukon River, and their trajectories tend to follow the general pattern of cyclonic circulation in the sound (i.e., Yukon River material enters the sound from the southwest, is transported north and northeast around the inside perimeter of the sound, and exits from the northwest). These data are supported by the salinity and temperature measurements, which indicated movements of low-salinity ($12\text{--}24\text{‰}$), relatively warm ($10\text{--}11\text{ °C}$) water to the northeast along the coast. These results are consistent with the general conclusions of Sharma et al. (1974) from suspended matter data obtained in August 1973. They are also

consistent with dispersal patterns of the Yukon River plume inferred from LANDSAT satellite photographs (Nelson et al. 1975). For example, Fig. 20-5 shows a LANDSAT photograph of the Yukon River plume taken on 20 July 1979. The plume, which appears in lighter grey tones than the less turbid water, can be traced as far north as approximately 70 km from the Yukon River Delta and as far east as 50 km from Stuart Island. These features are also consistent with the data of Cacchione and Drake (1979) from surveys made during quiescent periods in September 1976 and July 1977. Thus, it appears that the transport processes described above predominate throughout

Figure 20-5. MSS Band 5 of LANDSAT images E-21640-21360-5 and E-21640-21363-5 taken on 20 July 1979, showing evidence of transport of suspended matter (appearing lighter in tone than the less turbid water) into Norton Sound.

the region, at least during periods of calm weather in the summer.

The near-bottom distribution of salinity, temperature, and total suspended matter also gives evidence of cyclonic movement of low-salinity (20-22‰), warm (~10°C) water to the northeast along the coast. This water mass can be traced as far

north as Cape Darby. Near-bottom concentrations of suspended matter were highest near the mouth of the Yukon River and in the region about 20-30 km south-southwest of Nome. The near-bottom plume just seaward of the Yukon River extended to the northeast along the coast in a manner very similar to the surface plume. The near-bottom concentrations were

Figure 20-6. Vertical cross section of the distribution of: (a) salinity (b) temperature (c) sigma-t (d) total suspended matter for transect I in Norton Sound.

generally higher than surface concentrations, indicating that: (1) some fraction of the Yukon River material had settled to the near-bottom region during transit, and/or (2) a portion of the bottom sediments had been resuspended and remained in suspension.

Figs. 20-6 and 20-7 show cross sections of total suspended matter, salinity, temperature, and sigma-t for two east-west transects across the length of Norton Sound. Transect I is near the middle of the sound and Transect II is in the northern half, approximately 20-30 km from the coast. The two transects show very similar water mass characteristics. On the eastern side of the sound, both transects showed evidence of a two-layer system with the pycnocline varying in depth between 8 and 14 m (e.g., Stations 3 and 8 from Transect I and Stations 1, 9, and 10 from Transect II). Concentrations of suspended matter in this region showed a steady increase from about 6-8 m to the bottom. In the middle region (Stations 15, 20, and 29 of Transect I and Stations 22, 25, 27, and 39 of Transect II), the water column was virtually unstratified and concentrations of suspended matter increased to values greater than 30

mg/l in the near-bottom waters (e.g., Station 25 in Transect II). These water properties suggest that there is a tidally induced frontal zone with intense enough vertical mixing to break down the stability structure, with subsequent resuspension of sediments throughout the water column in some locations. This feature has the same characteristics as the anomalous plume of suspended matter described by Sharma et al. (1974) from data obtained in August 1973 at the same location. These data support the general conclusion of Cacchione and Drake (1979): during quiescent periods in the sound, resuspension of bottom sediments occurs as a result of increased tidal mixing during spring tide.

Further seaward, the water column was moderately stratified and concentrations of suspended matter decreased to values below 0.5 mg/l in surface waters. In near-bottom waters, however, concentrations of suspended matter were generally greater than 2.0 mg/l. The enriched concentrations of suspended matter near the bottom were probably caused by a combination of factors, including advective transport of particle-laden water to the northwest from

Figure 20-7 Vertical cross section of the distribution of: (a) salinity (b) temperature (c) sigma-t (d) total suspended matter for Transect II in Norton Sound.

within Norton Sound (Muench et al., Chapter 6, this volume) and local resuspension of bottom sediments.

Particulate elemental composition

In order to determine regional variations of the chemical composition of suspended material in Norton Sound, the particulate samples from the July 1979 cruise were analyzed for their major and trace element content by the methods described previously. The resulting data have been separated into five regions: Yukon River estuary with salinities of less than 15‰, Yukon River estuary with salinities between 15 and 25‰, eastern Norton Sound, central Norton Sound, and western Norton Sound/northeastern Bering Sea shelf. The averaged chemical data, along with published data for the Yukon River, are given in Tables 20-1 and 20-2. Table 20-3 shows C/N and element/Al ratios for the averaged data.

The elemental concentrations and elemental ratios illustrate some differences in composition be-

tween the suspended material discharged from the Yukon River and suspended matter in the sound. These differences can be viewed in terms of relative percentages of aluminosilicate and organic matter. Since most of the aluminum in marine particulate matter is in aluminosilicate material (Sackett and Arrhenius 1962), the Al concentrations in the suspended matter multiplied by 10 can be used to estimate percentages of aluminosilicate in the particulate matter. Similarly, Gordon (1970) suggested that particulate carbon multiplied by a factor of 1.8 may be used to estimate the amount of organic material in the suspended matter. On the basis of the particulate Al and particulate C concentrations, the composition of the suspended matter from the Yukon River estuary was determined to be approximately 88 percent aluminosilicate material and 6 percent organic matter. In like manner, samples from eastern and central Norton Sound are determined to contain about the same percentage of aluminosilicate material (88-92 percent). These results illustrate the predominance of the detrital

TABLE 20-1

Comparison of the elemental composition of suspended material from the Yukon River with the composition of suspended material collected from the near-shore regions seaward of the mouths of the Yukon River distributaries. (Samples collected 11 and 12 July 1979.¹)

Sample Description	No. of Samples	C ⁴ Wt. % ±1σ	N ⁴ Wt. % ±1σ	Mg Wt. % ±1σ	Al Wt. % ±1σ	Si Wt. % ±1σ	K Wt. % ±1σ	Ca Wt. % ±1σ	Ti Wt. % ±1σ	Cr ppm ±1σ	Mn ppm ±1σ	Fe Wt. % ±1σ	Ni ppm ±1σ	Cu ppm ±1σ	Zn ppm ±1σ
<i>Yukon River Suspended Material</i>															
Yukon River at Klakanak ²										147	1079	5.4	109	320	
Yukon River at Pilot Station ³		0.24-3.8								48	788-1308	3.1-4.3		24-148	49-142
<i>Yukon River Estuary</i>															
Surface Samples (0-15°/oo)	6	2.9 ±0.6	0.2 ±0.04	2.3 ±0.7	8.2 ±1.3	30.6 ± 1.9	2.2 ±0.3	1.5 ±0.2	0.50 ±0.06	110 ± 15	992 ± 131	5.5 ±0.8	59 ± 8	59 ± 8	171 ± 49
<i>Yukon River Estuary</i>															
Surface Samples (15-25°/oo)	6	4.2 ±1.2	0.4 ±0.8	3.1 ±0.8	9.3 ±1.8	31.5 ± 3.7	2.1 ±0.2	1.6 ±0.03	0.52 ±0.03	129 ± 15	1299 ± 192	5.81 ±0.4	60 ± 5	61 ± 10	193 ± 30

¹ Surface samples collected with precleaned polyethylene bottles. Standard deviations are given only for data obtained during a single sampling event wherever applicable.

² Data from Gibbs (1977)

³ Water Resources Data (1976-77) U.S. Geological Survey

⁴ Weight percentages of C and N were determined using two different filter types (Selas silver filters and Nucleopore filters) and, therefore, are subject to a greater number of errors than the data obtained for the inorganic elements, which were obtained from a single filter type.

TABLE 20-2
Summary of the elemental composition of suspended material collected from selected locations
in Norton Sound and northeastern Bering Sea shelf. (Samples were collected 7-18 July 1979.¹)

Sample Description	No. of Samples	C ² Wt. % ±1σ	N ² Wt. % ±1σ	Mg Wt. % ±1σ	Al Wt. % ±1σ	Si Wt. % ±1σ	K Wt. % ±1σ	Ca Wt. % ±1σ	Ti Wt. % ±1σ	Cr ppm ±1σ	Mn ppm ±1σ	Fe Wt. % ±1σ	Ni ppm ±1σ	Cu ppm ±1σ	Zn ppm ±1σ
<i>Eastern Norton Sound</i>															
Surface	7	15.4 ± 5.8	2.4 ± 0.9	3.1 ± 0.5	9.2 ± 1.4	30.1 ± 3.7	1.7 ± 0.3	1.4 ± 0.3	0.44 ± 0.06	199 ± 59	2346 ± 845	5.3 ± 0.6	52 ± 8	60 ± 9	201 ± 41
5 m above bottom	7	10.1 ± 6.2	1.1 ± 0.8	2.9 ± 0.5	9.1 ± 1.3	31.0 ± 2.7	2.0 ± 0.2	1.4 ± 0.2	0.53 ± 0.08	144 ± 38	2182 ± 682	5.7 ± 0.4	57 ± 5	62 ± 10	276 ± 289
<i>Central Norton Sound</i>															
Surface	18	14.6 ± 8.6	1.9 ± 1.1	3.3 ± 0.7	9.0 ± 1.7	30.8 ± 4.8	1.7 ± 0.4	1.7 ± 0.2	0.48 ± 0.08	148 ± 36	2672 ± 1104	5.4 ± 0.8	59 ± 21	61 ± 12	246 ± 90
5 m above bottom	18	5.6 ± 3.6	0.7 ± 0.5	3.0 ± 0.6	8.8 ± 1.2	32.4 ± 2.5	2.0 ± 0.2	1.6 ± 0.6	0.54 ± 0.06	138 ± 27	1797 ± 287	5.8 ± 0.6	57 ± 7	58 ± 6	196 ± 65
<i>Western Norton Sound/northeastern Bering Sea Shelf</i>															
Surface	18	25.6 ± 6.8	4.0 ± 0.9	0.9 ± 0.7	3.2 ± 1.4	20.0 ± 7.9	0.5 ± 0.2	1.4 ± 0.4	0.23 ± 0.10	100 ± 93	2160 ± 1392	2.3 ± 0.9	29 ± 17	50 ± 39	194 ± 111
5 m above bottom	18	12.3 ± 6.9	1.3 ± 0.4	1.9 ± 0.5	5.1 ± 1.2	31.8 ± 3.5	0.9 ± 0.3	1.2 ± 0.2	0.31 ± 0.07	81 ± 17	1506 ± 761	3.4 ± 0.7	30 ± 13	36 ± 11	137 ± 60

¹ Samples were collected with 10-l Niskin Bottles.

² Weight percentages of C and N were determined using two different filter types (Selas silver filters and Nuclepore filters) and, therefore, are subject to a greater number of errors than the data obtained for the inorganic elements, which were obtained from a single filter type.

TABLE 20-3
Average C/N and element/Al ratios for suspended materials
from the Yukon River estuary, Norton Sound, and northeastern Bering Sea Shelf

Sample Description	C/N	C/Al	N/Al	Mg/Al	Si/Al	K/Al	Ca/Al	Ti/Al	Cr/Al $\times 10^{-3}$	Mn/Al $\times 10^{-3}$	Fe/Al	Ni/Al $\times 10^{-3}$	Cu/Al $\times 10^{-3}$	Zn/Al $\times 10^{-3}$
Yukon River Estuary (0-15°/co)	14.5	0.35	0.02	0.28	3.73	0.27	0.18	0.06	1.34	12.1	0.66	0.72	0.72	2.08
Yukon River Estuary (15-25°/oo)	10.5	0.45	0.04	0.33	3.39	0.23	0.17	0.06	1.38	14.0	0.62	0.64	0.66	2.07
Eastern Norton Sound Surface	6.4	1.71	0.27	0.34	3.34	0.19	0.16	0.05	2.21	26.1	0.59	0.58	0.66	2.23
5 m above bottom	9.2	1.11	0.12	0.32	3.41	0.22	0.15	0.06	1.58	24.0	0.63	0.63	0.68	3.03
Central Norton Sound Surface	7.7	1.62	0.21	0.37	3.42	0.19	0.19	0.05	1.64	29.7	0.60	0.65	0.67	2.73
5 m above bottom	8.0	0.63	0.08	0.34	3.68	0.23	0.18	0.06	1.57	20.4	0.66	0.65	0.66	2.23
Western Norton Sound-northeastern Bering Sea shelf														
Surface	6.4	8.0	1.30	0.28	6.35	0.16	0.44	0.07	3.12	67.5	0.72	0.91	1.56	6.06
5 m above bottom	9.5	2.4	0.25	0.37	6.23	0.17	0.24	0.06	1.59	29.5	0.66	0.59	0.67	2.68

material from the Yukon River in the central and eastern regions of the sound. This finding is supported by the chemical data for Si, K, Ca, Ti, Fe, Ni, and Cu, which are at about the same levels of concentration in eastern and central Norton Sound as they are in the Yukon River estuary. Only C, N, Mn, and Zn show enrichments offshore. For C and N these enrichments are attributed to a relative increase in the concentration of marine organic matter in offshore waters, which is probably due to increased light penetration away from the zone of high turbidity. This conclusion is supported by the C/N ratios (Table 20-3), which show a general decrease seaward, indicating a transition from organic matter dominated by terrestrial material (C/N ratios ranging between 12 and 15) to organic matter dominated by material of marine origin (C/N ratios ranging between 6 and 9) (Loder and Hood 1972). Mn and Zn enrichments can be attributed to a number of processes which are discussed in detail later.

In the western Norton Sound/northeastern Bering Sea shelf region, the suspended matter was depleted in particulate Mg, Al, K, Ti, Fe, Ni, and Cu and enriched in particulate C and N relative to the Yukon River estuarine samples. The depletions are attributed to a drop in the relative amount of aluminosilicate material in the suspended matter (<52 percent by weight) and an increase in the proportion of marine organic matter (>40 percent by weight), which contains less Mg, Al, K, Ti, Mn, Fe, Ni, and Cu than aluminosilicate material (Martin and Knauer

1973). It is important to note, however, that on the average the samples from this region contain about 88 percent more Mn than the Yukon River estuarine samples. Similarly, Zn concentrations in the suspended matter from this region are about the same as in the estuarine samples even though there is a significant drop in relative amount of aluminosilicate material in the suspended matter. These findings indicate that Mn and Zn concentrations in the suspended matter are controlled by distinctly different chemical processes.

In an attempt to determine the chemical nature and source of the enriched Mn and Zn in the offshore suspended matter, selected surface and near-bottom samples were treated with 25 percent (v/v) acetic acid to separate poorly structured oxyhydroxides from the more crystalline phases. This procedure has been shown to selectively dissolve trace elements precipitated in acid-soluble metal oxides and those adsorbed onto mineral surfaces without affecting highly oxidized ferromanganese minerals or the lattice structure of clays (Hirst and Nicholls 1958, Chester and Hughes 1967, and Bolger et al. 1978). The results of these experiments are given in Table 20-4. The data show higher amounts of weak-acid-soluble Mn in the offshore samples than in the estuarine samples, which are significant at the $p < 0.05$ level. These higher amounts, computed by taking the differences between the offshore and estuarine samples as a ratio to the estuarine samples, range between 134 percent and 351 percent and

TABLE 20-4
Partitioning of Mn and Zn between weak-acid-soluble (WAS)
and weak-acid-insoluble (WAI) fractions of suspended material
from Norton Sound and northeastern Bering Sea Shelf
(Data presented as a percentage of total suspended matter)

Sample location	No. of samples	WAS Mn $\pm 1\sigma$	WAI Mn $\pm 1\sigma$	WAS Zn $\pm 1\sigma$	WAI Zn $\pm 1\sigma$
Yukon River Estuary	3	0.066 ± 0.017	0.052 ± 0.006	0.0059 ± 0.0028	0.0140 ± 0.0031
Eastern Norton Sound	5	0.155 ± 0.038	0.040 ± 0.011	0.0095 ± 0.0031	0.0099 ± 0.0021
Central Norton Sound	9	0.184 ± 0.085	0.054 ± 0.017	0.0108 ± 0.0056	0.0114 ± 0.0026
Western Norton Sound	9	0.298 ± 0.092	0.074 ± 0.041	0.0107 ± 0.0053	0.0125 ± 0.0070

account for all of the excess Mn in the suspended matter. Similarly, the data for Zn in the weak-acid-soluble fraction show enrichments ranging between 61 percent and 83 percent in the offshore samples which are significant at the $p < 0.20$ level. These results indicate that in the offshore waters Mn and Zn are being concentrated in the weak-acid-soluble fraction of the particulate matter, which in these samples probably consists of poorly structured oxyhydroxides of Mn. The probable sources of this material will be discussed below.

DISCUSSION

Probable source of excess Mn in the suspended matter

There are several possible sources of the excess Mn in the suspended matter of Norton Sound. These include (1) differential settling of particles of various sizes; (2) resuspension of Mn-enriched sediments; and (3) reductive dissolution of Mn within recent sediments followed by oxidative precipitation of Mn onto particulate phases in the water column. The first mechanism is unlikely in view of Gibbs's (1977) data for the chemical variations in the various size-fractions of Yukon River suspended material. The mean particle size distribution of suspended material in the sound would have to be about an order of magnitude smaller (i.e., a decrease from an average size of about 20 μm to about 2 μm) for the two-to-threefold increases in total Mn to occur. Unless

some unusual chemical interactions were occurring in the estuary, this would necessarily be accompanied by a similar enrichment of total Fe and Cu in the suspended matter. No enrichments of that magnitude were observed in the Fe and Cu data. Furthermore, the particle-size data of Cacchione and Drake (1979) indicate that suspended matter in Norton Sound is primarily composed of fine-to-medium silt in the range between 4 and 32 μm . These data indicate that if differential settling occurs in Norton Sound, it is definitely not of the magnitude required to produce the observed Mn enrichments in the suspended matter.

The resuspension mechanism can also be refuted using a similar argument. While the distributions of suspended matter indicated that bottom sediments were being resuspended, the Mn content of the bulk sediments has been reported to be only in the range of 600-1,650 ppm (Larsen et al. in press). This means that the Mn content of the resuspended material would have to exceed the concentration observed within the sediments by a factor of about 2-4 to account for the observed Mn concentrations in the suspended matter. This would occur only if the clay size-fraction of the sediments were being preferentially resuspended. Since the particle-size data of Cacchione and Drake (1979) do not show any evidence for a decrease of this kind, this mechanism does not seem likely.

Reduction of Mn after burial in recent sediments with accompanying upward transport of dissolved

Mn into the overlying water, followed by precipitation onto suspended matter best explains the observed data. Efflux of Mn from rapidly accumulating sediments has been reported for several estuarine and coastal environments (Elderfield 1976, Graham et al. 1976, Aller 1977, Trefry 1977, Yeats et al. 1979, and Massoth et al. 1979). From studies of the sediments extending seaward of the Mississippi River, Trefry (1977) found that Mn fluxes from recent sediments varied directly with sedimentation rate. High Mn fluxes (i.e., $\sim 2.7 \mu\text{g}/\text{cm}^2/\text{d}$) were observed in sediments that accumulate at a rate of about $2.0 \text{ g}/\text{cm}^2/\text{yr}$, whereas low Mn fluxes ($0.71 \mu\text{g}/\text{cm}^2/\text{d}$) were observed in sediments that accumulate at a rate of $0.08 \text{ g}/\text{cm}^2/\text{yr}$. In Norton Sound modern sediments with accumulation rates ranging from 0.05 to $0.17 \text{ g}/\text{cm}^2/\text{yr}$ cover an area of approximately $22,000 \text{ km}^2$ (Nelson and Creager 1977). Assuming an average sedimentation rate of $0.1 \text{ g}/\text{cm}^2/\text{yr}$ for these sediments and using linear interpolation of Trefry's (1977) Mn flux data (i.e., $0.68 \mu\text{g}/\text{cm}^2/\text{d}$), approximately $1.5 \times 10^8 \text{ g}$ Mn would be released daily into Norton Sound from this source. At this rate it would require approximately 21 days to account for all of the estimated excess Mn in the particulate matter (approximately $3.1 \times 10^9 \text{ g}$ Mn, assuming a total area of $45,000 \text{ km}^2$, an average depth of 16 m, an average concentration of suspended matter of $4.0 \text{ mg}/\text{l}$, and an average concentration of excess Mn of $1,079 \text{ ppm}$). If it is assumed that the rate of Mn oxidation is fast relative to an accumulation time of 21 days, then contact periods approximately equal to this time would be required for the chemical interactions to occur. Circulation in the sound is not completely understood, but studies conducted in summer indicate relatively sluggish circulation (Muench et al., Chapter 6, this volume). Net currents, with speeds varying between 10 and $15 \text{ km}/\text{d}$ in surface waters and between 1 and $4 \text{ km}/\text{d}$ in deep water, have been measured for short periods of time. Using a mean current of $8 \text{ km}/\text{d}$ and a mean travel distance of 400 km , it is estimated that about 50 days are required for water to pass through the sound—a little more than twice the time required for the Mn from the sediment to accumulate on the suspended matter. Thus, if the underlying assumption that the kinetic rate of Mn oxidation in coastal waters is relatively rapid is correct, then the sediments could easily be the major source of the excess Mn in the suspended matter. The assumption of a rapid rate for Mn oxidation is supported by the recent findings of Wollast et al. (1979), to the effect that Mn oxidation in the Rhine and Scheldt estuaries is essentially complete within

10 days and the process is mediated by several strains of marine bacteria indigenous to coastal environments.

Implication for the geochemistry of Zn in Norton Sound

This discussion of the geochemical behavior of Mn in the sound is also important for understanding the chemical behavior of Zn in the suspended matter. As noted earlier, both Zn and Mn are enriched in the weak-acid-soluble fraction of the particulate matter. This is probably due to adsorption and/or coprecipitation of Zn on or in the newly formed Mn oxyhydroxides. Fig. 20-8 shows a plot of the relationship between total Zn and total Mn and for both surface and near-bottom samples. The plot of total Zn versus total Mn is roughly linear ($r = 0.60$), indicating an association between these two metals in the particulate matter. These results suggest that as the Mn oxyhydroxides form on the particulate matter, Zn is scavenged from solution. In similar fashion, the relationship between weak-acid-soluble Zn and weak-acid-soluble Mn is also linear ($r = 0.39$). This process effectively concentrates Zn and Mn in the suspended matter, which eventually either settles to the bottom of the sound or is transported to the northwest into the northeastern Bering Sea shelf and beyond.

Figure 20-8. Plot of the relationships between total particulate Zn vs. total particulate Mn and weak-acid-soluble Zn vs. weak-acid-soluble Mn for selected surface and near-bottom samples from Norton Sound and northeastern Bering Sea shelf.

Implications for the geochemistry of suspended materials and sediments in regions beyond Norton Sound

The physical, chemical, and biological processes affecting the distribution and chemical composition

of suspended matter in Norton Sound and the northeastern Bering Sea shelf may also affect the composition of suspended materials and sediments in oceanic waters beyond the Bering Sea. As we stated earlier, the work of Nelson and Creager (1977) indicates that as much as one-third of the sediment load of the Yukon River bypasses the northern Bering Sea to accumulate in a thick deposit of Holocene sediment in the southern Chukchi Sea. These authors suggest that resuspension and transport of previously deposited sediments from Norton Sound during storms may contribute significantly to this phenomenon. Further support for this mechanism is indicated by the recent findings of Cacchione and Drake (1979) that more than 50 percent of the sediment transport in Norton Sound occurs during storm activity. Since the resuspended sediments in Norton Sound become enriched in Mn and Zn in oxygenated waters, it is reasonable to expect that the particulate matter will remain enriched in these metals until the particles are buried within the sediments and the recycling process is reinitiated in the reducing zone (Elderfield 1976). If it is further assumed, as suggested by Yeats et al. (1979), that the Mn precipitating in the water column will tend to be preferentially associated with small-sized particles, then the Mn and Zn content of the suspended material could be further enriched if some of the larger particles settle out. Therefore, Mn and Zn could be continually enriched in the particulate matter that is transported past the Bering Sea into the Chukchi Sea, where it forms a major fraction of the suspended matter and the recent sedimentary deposits. Thus, Mn and possible Zn recycling in the Chukchi Sea may be even more pronounced than in Norton Sound, because the incoming particulate materials are significantly enriched in these metals and the accumulation rates for recent sediments in the Chukchi Sea are as great as or greater than in Norton Sound (Naidu and Sharma 1972, Nelson and Creager 1977). If this is true, then Mn and Zn can undergo a number of recycling processes before they are ultimately buried in continental shelf, slope, or deep ocean sediments.

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Some Heavy Metal Contents of Bering Sea Seals

21

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ABSTRACT

The Cd, Cu, Ni, and Zn contents of liver, kidney, and muscle tissue of a suite of seal samples taken from the central Bering Sea are given. Heavy metals, particularly cadmium, are highest in livers and kidneys. Spotted seals have the lowest contents of heavy metals, and, tentatively, bearded seals the highest. However, no clear correlation between the contents of these metals and feeding habits or age could be discerned from the data base.

INTRODUCTION

This chapter presents baseline data obtained as part of the BLM/NOAA Outer Continental Shelf Environmental Assessment Program for Alaska. The samples were originally collected for other purposes. Nevertheless, we have obtained a number of precisely analyzed heavy metal data for an area—the central Bering Sea—which is of considerable economic importance and current oceanographic interest, but for which very little chemical characterization exists.

The major portion of this work concerns heavy metal contents of various seal tissue samples. These distributions are considered in the context of the environmental baselines determined for water and sediments within coincident and adjacent portions of the Bering Sea.

ANALYSIS TECHNIQUES

All biota and sediment samples considered here have been analyzed spectrophotometrically for selected heavy metals. The major limitations on analytical precision for both classes of samples are connected with the dissolution step and inter-element matrix effects. The problem of oxidizing the tissue samples without loss of volatile metals was addressed here by low-temperature ashing in an oxygen plasma

furnace. The residues from this step were then treated with Ultrex nitric acid in a teflon digestion bomb prior to graphite furnace atomic spectrometric analysis. Standard curves run with each batch were prepared by adding standards to a matrix prepared in bulk for each type of tissue sample. NBS standards were carried through with each batch to monitor accuracy.

Soluble seawater trace metal data are only a very minor component of this particular report. The stringent sampling strategy needed has been discussed in detail by Burrell (1978). Final analysis was by differential pulse anodic stripping voltammetry.

ENVIRONMENTAL DATA

By means of the careful sampling procedures described by Burrell (1978; Heggie, unpublished data), a number of water samples were collected from the U.S.S.R. hydromet vessel *Volna* on the stations shown in Fig. 21-1 in July-August 1977. Coincidentally, many of these stations are in that part of the Alaskan shelf west of Nunivak Island from which the seal specimens discussed below were collected the previous spring.

Soluble copper and lead values have been determined and Fig. 21-2 shows the mean ranges for over 100 samples collected through the water column (Heggie, unpublished data). These closely conform to distributions given previously for the Gulf of Alaska and elsewhere. Since mean soluble concentrations vary only between narrow limits in the open ocean, this similarity was expected. However, nanogram ranges have been suggested only recently for these metals; our data support such levels for

Figure 21-1. Stations occupied by U.S.S.R. hydromet vessel *Volna* July-August 1977.

unpolluted open-ocean water and at the same time add confidence to their accuracy. Although we have good data on the soluble portions of copper and lead only, there is no evidence of any anomalous regional trends, and it is expected that the concentrations of other trace metals in solution will show the same low ranges given in earlier publications for the Gulf of Alaska (e.g., Burrell 1978).

The geochemistry of the surface sediments is discussed in Chapter 19 and need not be considered further here.

HEAVY METAL CONTENTS OF MARINE MAMMALS

Fig. 21-3 shows the localities of sacrificed seal samples collected on two separate cruises in March-April and May-June of 1977.

The original objective of this project was to look for statistical differences between the heavy metal contents of four species of seal which were thought to have distinctive feeding habits. Our data are

Figure 21-2. Mean ranges for soluble ($<0.4 \mu\text{m}$) and lead concentrations for samples collected from the surface and adjacent to the bottom at the stations given in Fig. 21-1 (D. T. Heggie, unpublished data).

Figure 21-3. Localities of seal samples collected on O.S.S. *Surveyor* cruise March-April 1977 (open symbols) and O.S.S. *Discoverer* cruise May-June 1977 (closed symbols).

largely for ribbon, spotted, and bearded seals (Tables 21-1 and 21-2), which were believed to feed predominantly on demersal fish and benthos, pelagic fish, and invertebrate benthos respectively. Unfortunately it was not found feasible to obtain representative food species when the mammals were collected. Nor were stomach contents suitable for either identification or analysis: in many cases it was found that the animals had starved for a number of days. The biological investigators also found

the seals to be largely opportunistic feeders, so that diet differentiation by species, where it occurred, was observed only where there was adequate choice.

For most of the individuals collected we have analyzed for cadmium, nickel, copper, and zinc in muscle, liver, and kidney tissue and these data, as means of duplicate determinations, are given in Tables 21-3 and 21-4. Table 21-5 gives accuracy and precision information relating to this batch of numbers.

TABLE 21-1

Bering Sea
O.S.S. *Surveyor* 31 March-27 April 1977
Seal samples collected for heavy metal analysis

Sample #	Latitude (N)	Longitude (W)	Species	Sex	Weight (kg)	Age (yrs)
1	58°51.0	173°08.0	Ribbon	F	39.5	1
2	58°51.0	173°08.0	Ribbon	M	102.0	—
3	58°56.0	172°40.0	Ribbon	M	81.8	4
4	58°45.6	172°55.4	Spotted	M	35.0	1
5	59°00.6	173°15.0	Bearded	F	181	6+
6	58°53.0	173°07.0	Ribbon	M	107.3	15
7	58°43.9	169°32.9	Bearded	M	232	12
8	58°48.5	169°41.0	Bearded	F	227	12+
10	59°06.3	169°41.3	Spotted	F	41.8	1
11	58°24.7	164°52.3	Ribbon	F	98.6	3
16	58°21.3	164°49.7	Bearded	F	204.5	2
28	58°54.2	169°13.6	Spotted	M	89.9	6
29	58°40.1	169°40.3	Ribbon	M	59.9	3
30	58°34.8	169°28.8	Spotted	M	84.0	7
32	59°22.5	173°43.0	Spotted	M	118.0	17

As noted above, marked differences in heavy metal contents as a reflection of a particular type of food would not be expected. Nevertheless, Table 21-6 lists some possible trends based on this very limited sample batch. It appears likely that the spotted seals have the lowest metal contents and—even more tentatively—bearded seals the highest. The former are considered to be largely consumers of finfish, whereas the bearded seals eat large quantities of benthic invertebrates. Since more metals are concentrated in the benthos than in pelagic communities, this possible trend is of interest. Clearly more positive correlation would require not only a considerably larger number of specimens of one species but also concurrent food species or, preferably, fresh stomach contents.

Liver and kidney contents show generally elevated levels of these metals, as would be expected. Cadmium in these organs is notably high, but the general lack of comparable reference data does not permit comment as to whether it is unusually so. Olafson and Thompson (1974) have reported on the isolation of metallothioneins from seal livers, and have suggested that the biosynthesis of such cadmium-binding protein acts primarily as a detoxification mechanism in these, as in terrestrial mammals. Assuming that

these complexes serve no other metabolic function, then their presence implies toxic ambient marine levels of these metals (notably cadmium, but also mercury and zinc).

The mean kidney content of cadmium for all the analysis samples is around 24 ppm dry weight: a "critical" liver concentration in man has been estimated at around 200 ppm. Kerfoot and Jacobs (1976) have echoed earlier models in which a daily intake of some 50 $\mu\text{g/day}$ of cadmium will produce the observed mean body burden of around 30 mg, which would approximately correspond to a critical organ concentration of around 50 ppm. Estimates such as these assume a long residence time for cadmium in the mammalian tissue: this is one of the chief health hazards of this metal. There appears to be no correlation of liver and kidney contents of metals with the age of the animals given in this report, however. In the muscle tissue, contents of these metals, especially zinc, may increase with age; but, in general, the assumption of relatively short residence times with varying organ contents reflecting recent eating habits is attractive. (Note also that contents of cadmium in livers, for example, of the May samples are generally lower than in samples taken in April.) We have here, however, far too few individuals of any one species or age group to permit anything approaching a rigorous statistical analysis.

TABLE 21-2

Bering Sea
O.S.S. *Discoverer* 25 May-June 1977
Seal samples collected for heavy metal analysis

Sample #	Latitude (N)	Longitude (W)	Species	Sex	Weight (kg)	Age (yrs)
1	60°37.7	174°27.2	Ribbon	M	80.0	10
2	60°36.3	174°37.5	Spotted	F	57.7	10
3	60°36.3	174°37.5	Spotted	M	45.5	0.3
5	60°36.3	174°37.5	Spotted	M	49.1	2
7			Ribbon	M	73.6	7
8	60°26.5	168°55.8	Spotted	F	55.0	4
10	60°24.2	169°49.8	Spotted	M	68.6	5
13	60°35.9	168°10.1	Ringed	M	8.7	1
15	60°56.6	170°48.3	Spotted	F	37.7	<1

TABLE 21-3
Bering Sea
O.S.S. Surveyor 31 March-27 April 1977
Heavy metal contents of seal tissue ($\mu\text{g/g}$ dry weight)

Sample	Species	Tissue	Cd(a)			Ni(a)			Cu(a)			Zn(a)		
01	Rib	muscle	0.13	±	0.01	2.5	±	0.4	6.7	±	0.5	37	±	12
		liver	6.4	±	0.1	2.6	±	0.9	16.5	±	0.2	169	±	7
		kidney	53.0	±	0.3	2.8	±	0.2	19.4	±	0.1	149	±	1
02	Rib	muscle	0.3		(c)	1.3		(c)	4.4	±	0.6	50	±	0
		liver	8.7	±	1.2	1.3		(c)	16.5	±	3.5	8	±	2
		kidney	20.1	±	4.5	0.5	±	0	16.5	±	3.5	102	±	22
03	Rib	muscle	0.25	±	0	8.3	±	1.0	5.5	±	0	54	±	2
		liver	6.4	±	0.1	2.2	±	0.4	26.9	±	1.3	140	±	15
		kidney	34.5		(c)	3.0		(c)	16.0		(c)	98		(c)
04	S	muscle	0.14	±	0.01	3.0	±	0.2	6.7	±	1.3	51	±	9
		liver	0.4	±	0.1	2.5	±	0.2	25.0	±	2.5	116	±	51
		kidney	16.7	±	0.2	2.4	±	0.4	44	±	13	113	±	23
05	B	muscle	0.8	±	0.1	2.8	±	0.3	7	±	2	175	±	25
		liver	21.0	±	0.5	1.3	±	0.1	36.5		(c)	182	±	2
		kidney	16.0	±	2.0	2.0	±	0.1	28.5	±	0	163	±	3
06	Rib	muscle	0.47	±	0.01	5.8	±	0	6.8	±	0.8	18	±	6
		liver	11.4	±	0.1	0.9	±	0.2	29.0	±	2.5	100	±	10
		kidney	33.4	±	0	1.2	±	0.2	17.5	±	0.3	48	±	0
07	B	muscle	0.57	±	0.04	2.8	±	0.2	<5		(b)	147	±	20
		liver	22.2	±	1.3	0.8	±	0	22.8	±	0.2	170	±	17
		kidney	22.5	±	2.0	1.3	±	0.3	40	±	6	160	±	15
08	B	muscle	1.26	±	0	1.2	±	0.1	7.6	±	0	40	±	1
		liver	41.1	±	0.9	0.5	±	0	44.1	±	2.4	87	±	3
		kidney	17.4	±	0.1	1.5	±	0.1	28.1	±	0.6	110	±	1
10	S	muscle	1.5	±	1.0	2.4	±	0.2	11	±	3	52.5	±	0
		liver	0.6	±	0.1	8.1	±	1.6	18	±	4	213	±	7
		kidney	44.3	±	0.7	<1.0		(b)	24.5		(c)	183		(c)
11	Rib	muscle	0.24	±	0.02	<0.5		(b)	16.5	±	3.5	140	±	5
		liver	35	±	15	<0.5		(b)	13.8	±	1.2	15	±	10
		kidney	16		(c)	<0.5		(b)	19	±	6	8	±	3
16	B	muscle	1.13	±	0.07	4.8	±	0.2	6.9	±	0.6	60	±	10
		liver	5.3	±	0.6	5.4	±	1.1	37.6	±	1.6	228	±	3
		kidney	108.7	±	0.5	6.2	±	0.2	34.7	±	0.2	110	±	20
17	Walrus	muscle	1.50	±	0.25	2.4	±	0.5	6.3	±	1.3	43	±	8
		liver	26.0	±	1.0	1.6	±	0.3	41.6	±	0.4	99	±	1
		kidney	26.5	±	1.0	<1.0		(b)	27.7	±	0	96	±	1
28	S	muscle	0.11	±	0.01	<0.5		(b)	6.4	±	0.05	68	±	1
		liver	4.5	±	0.05	<0.5		(b)	16.4	±	0.1	136	±	1
		kidney	—		—	—		—	—		—	—		—
29	Rib	muscle	0.70	±	0.03	<0.5		(b)	1.7	±	0	74	±	2
		liver	17.0	±	0.3	<0.5		(b)	5.2	±	1.4	130	±	3
		kidney	37	±	5	<0.5		(b)	5.2	±	1.3	92	±	14
30	S	muscle	0.16	±	0.04	<0.5		(b)	4.0	±	0.1	133	±	0
		liver	7.3	±	0.2	<0.5		(b)	18.5	±	0.2	160	±	0
		kidney	34	±	5	<0.5		(b)	15	±	2.5	120	±	17
32	S	muscle	0.17	±	0.01	<0.5		(b)	1.1	±	0.1	62	±	9
		liver	15.9	±	0.8	<0.5		(b)	17.1	±	0.9	125	±	23
		kidney	49	±	7	<0.5		(b)	4	±	2.5	110	±	32

(a) mean of duplicate determinations

(b) duplicate determinations

(c) single determinations

TABLE 21-4

Bering Sea
O.S.S *Discoverer* 25 May-June 1977
Heavy metal contents of seal tissue ($\mu\text{g/g}$ dry weight)

Sample	Species	Tissue	Cd(a)		Ni(a)		Cu(a)		Zn(a)	
01	Rib	muscle liver kidney	0.29 3.2 <0.3	$\pm$ 0.04 $\pm$ 0.9 (b)	0.5 <0.5 1.3	$\pm$ 0.1 (b) $\pm$ 0.5	4.6 6 10	$\pm$ 0.1 $\pm$ 3 $\pm$ 1	96 185 197	$\pm$ 5 $\pm$ 15 $\pm$ 6
02	S	muscle liver kidney	1.3 2.48 <0.5	(c) $\pm$ 0.02 (c)	<0.5 <0.5 <0.5	(b) (b) (b)	3.9 11 25.5	$\pm$ 0.5 $\pm$ 3 $\pm$ 0	18.2 23.3 22	$\pm$ 1.1 $\pm$ 0.2 $\pm$ 2
03	S	muscle liver kidney	0.29 1.85 11.0	$\pm$ 0.07 $\pm$ 0.15 (c)	2 2.5 1.4	$\pm$ 1 $\pm$ 0 (c)	7.6 16.5 24.0	$\pm$ 0.8 $\pm$ 0.1 (c)	87 114 93	$\pm$ 7 $\pm$ 1 $\pm$ 25
05	S	muscle liver kidney	0.6 0.99 <0.5	(c) $\pm$ 0.01 (c)	<0.5 <0.5 <0.5	(b) (b) (b)	2.2 2.3 23	$\pm$ 0.7 $\pm$ 0.1 $\pm$ 3	7.5 10 13	$\pm$ 1 $\pm$ 1 $\pm$ 1
08	S	muscle liver kidney	0.08 1.9 1.52	$\pm$ 0 $\pm$ 0.4 $\pm$ 0.02	<0.5 <0.5 <0.5	(b) (b) (b)	6.5 44.3 29	$\pm$ 0.5 $\pm$ 0.3 $\pm$ 3	62 101 83	$\pm$ 3 $\pm$ 15 $\pm$ 3
10	S	muscle liver kidney	0.24 2.6 11.0	$\pm$ 0 $\pm$ 0.4 $\pm$ 4	2.4 1.8 0.9	$\pm$ 0 $\pm$ 0.2 $\pm$ 0.1	7.1 20.3 12	$\pm$ 0.9 $\pm$ 0 $\pm$ 3	123 108 103	$\pm$ 2 $\pm$ 5 $\pm$ 9
11	Ring	muscle liver kidney	0.19 5 2.52	$\pm$ 0.01 (c) $\pm$ 0.04	0.6 <0.5 0.7	$\pm$ 0.1 (b) $\pm$ 0.1	9.5 25 10	$\pm$ 0 (c) (c)	135 72 227	$\pm$ 0 $\pm$ 2 $\pm$ 13
15	S	muscle liver kidney	0.17 3.1 13.1	$\pm$ 0.05 $\pm$ 0.4 $\pm$ 0.1	1.4 2.1 3.8	$\pm$ 0.8 $\pm$ 0.2 $\pm$ 0.3	6.9 18.5 16.1	$\pm$ 0.1 $\pm$ 1.5 $\pm$ 0.1	80 134 123	$\pm$ 1 $\pm$ 9 $\pm$ 1
07	Rib	muscle liver kidney	0.27 3 5.5	$\pm$ 0.01 $\pm$ 0.2 $\pm$ 0	<0.5 17.5 <0.5	(b) $\pm$ 0 (b)	3.3 2.5 10	$\pm$ 0.7 (c) $\pm$ 1	116 115 138	$\pm$ 7 $\pm$ 5 $\pm$ 3

(a) mean of duplicate determinations
(b) duplicate determinations
(c) single determinations

TABLE 21-5

Bering Sea
Marine mammals analysis program—precision and accuracy
($\mu\text{g/g}$ dry weight $\pm$ one standard deviation)

a. NBS Standard # 1571 orchard leaves

Element	n	This study	NBS certified
Cd	3	0.15 $\pm$ 0.06	0.11 $\pm$ 0.02
Ni	4	1.4 $\pm$ 0.03	1.3 $\pm$ 0.2
Cu	7	10.5 $\pm$ 3	12 $\pm$ 1
Zn	7	25 $\pm$ 5	25 $\pm$ 3

b. NBS Standard # 1577 bovine liver

Element	n	This study	NBS certified
Cd	3	0.31 $\pm$ 0.07	0.27 $\pm$ 0.04
Ni	4	0.9 $\pm$ 0.08	—
Cu	3	150 $\pm$ 30	193 $\pm$ 10
Zn	8	120 $\pm$ 20	130 $\pm$ 10

TABLE 21-6

Heavy metal distributions in seal tissue
from Bering Sea—Spring 1977

Cadmium

1. Concentrations in the kidneys of ribbon seals greater than in spotted or bearded
2. Higher cadmium contents in the muscle tissue of bearded seals than in ribbon or spotted
3. Liver contents of spotted seals relatively low
4. Liver contents of bearded seals relatively high

Nickel

Muscle contents generally higher than liver or kidneys

Copper

Spotted seal kidneys generally higher than liver, but reverse trend for ribbon and bearded seals

Zinc

1. Muscle contents of bearded (and possibly spotted) seals higher than ribbon
2. Concentration of zinc higher in livers than in kidneys of all species

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Preliminary Observations of the Carbon Budget of the Eastern Bering Sea Shelf

22

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ABSTRACT

Preliminary studies of the CO_2 system of the PROBES area of the Bering Sea shelf in May of 1976 and May and June of 1978 have shown partial pressures of CO_2 in the surface water as much as $250 \mu\text{A}$ (ppm) less than overlying air, which averaged about $329 \mu\text{A}$ (ppm). These low pressures are evidently produced by photosynthesis, but their persistence long after nutrients are depleted and photosynthesis is low indicates a large sink for CO_2 under earlier bloom conditions accompanied by limited respiration in the water column which would recycle the fixed organic carbon back to the inorganic carbon pool. If transfer of CO_2 through the sea surface from the atmosphere is on the order of grams of carbon per day, many months would be required to reach equilibrium without the help of respiration, which apparently occurs later in the season than the field work to date.

A high correlation between deficiency of CO_2 in the water with respect to air ($-\Delta p\text{CO}_2$) and nitrate was found when values of CO_2 were greater than $-130 \mu\text{A}$. When CO_2 was below $-130 \mu\text{A}$, nitrate tended to be zero while $p\text{CO}_2$ was found as low as $-230 \mu\text{A}$.

Total CO_2 measurements made during the phytoplankton bloom period (May-June 1978) gave values significantly lower ($1.5\text{--}1.8 \text{ mM}$) than those found under non-bloom conditions (2.00 mM). The CO_2 represented by the difference in total carbon dioxide between bloom and non-bloom conditions is apparently being held in the system as fixed carbon, either in the form of detritus available for consumption, or as the portion of primary production which is stored in the body tissues of flora and fauna.

INTRODUCTION

It is well known from many previous observations (Park et al. 1974, Gordon et al. 1973, Kelley et al. 1971, Park et al. 1958) that the $p\text{CO}_2$ of euphotic regions of the ocean becomes depressed with respect to the overlying air under conditions of high primary productivity. Although photosynthesis is the only known sink for molecular carbon dioxide, the carbon dioxide respired by mammals and birds directly into the atmosphere is an effective sink which may be of importance in some regions.

The supply of CO_2 to the euphotic zone during the season of maximum primary productivity is derived primarily from flux through the sea surface from the

overlying air, although additions from vertical transport from deeper water, respiration, and horizontal convection may contribute heavily at other seasons and in other circumstances. It is clearly established, however, that under positive net carbon fixation, stability in the water column structure conducive to phytoplankton growth limits vertical transport; and transport of properties by horizontal convection becomes negligible if surrounding waters are similar to the study area. Thus, under conditions of high primary productivity, the product of the difference in $p\text{CO}_2$ (in μA) between air and surface water and the exchange rate of CO_2 through the surface is directly related to the net photosynthesis (see equation 1). In the carbon budget net photosynthesis is represented by the sum of the flux of CO_2 through the sea surface and the decrease of total inorganic CO_2 in the water column, corrected for the direct precipitation of carbonate salts (e.g., carbonate exoskeleton formation).

If there is vertical transport, as in regions of upwelling, the stability of the water column is disrupted and the deep water rich in carbon dioxide reaches the surface. This water, which has been enriched with molecular carbon dioxide through biological processes at depth, is supersaturated with respect to air, and a net flux of CO_2 from the water to the air occurs. Hood and Kelley (1976) have used a measure of this flux to estimate vertical transport.

Carbon dioxide, a fundamental component in all metabolic processes, is closely coupled to many chemical and physical events in the ocean. It is easy to monitor the partial pressure of carbon dioxide in the surface water continuously at sea; this is a powerful qualitative tool for mapping the ocean for such features as upwelling, biological activity, currents, and freshwater influx. Moreover, since the concentration of all the components of the carbon di-

oxide system in sea water is sensitive to most oceanographic events, a detailed understanding of carbon dioxide dynamics in the environment will provide additional signals useful in evaluating the initiation, path followed, and fate of some otherwise poorly understood elements of the ecosystem.

This study of the carbon budget was undertaken in the eastern Bering Sea to assist the PROBES (Processes and Resources of the Bering Sea Shelf) project in understanding biological energy flow and trophic dynamics in the important biological resource area often known as the "golden triangle." The base of this triangular region extends from just north of Unimak Pass running east to the 80-m bathymetric contour. The two sides join near St. George Island, forming the triangle.

The PROBES project seeks to follow the sequence of events involved in energy flow and efficiency of food utilization in the biota at the lower trophic levels of the biological community by following the early life cycle of the walleye pollock (*Theragra chalcogramma*), in the hope that it will function as a tracer of other organisms in this trophic level. This study requires the detection of the early changes in the system that indicate the conditions for onset of zooplankton spawning, phytoplankton productivity, and the hatching of the widely scattered pollock eggs (see chapters by Hattori and Goering, Cooney, and Nishiyama, in Volume 2 of this book). More conventional methods for following ecosystem changes that affect the development of the biological community—monitoring temperature, light, nutrient regime, water-column structure, and the presence of biological community components—may not be adequate to define the fine structure on which these processes depend for initiation. Data on the changes in the carbon dioxide system that occur during the course of ecosystem development may provide additional information needed to explain how the system functions.

The carbon dioxide cycle

To do budgetary studies of carbon in the marine environment it is first necessary to determine the pathways and fates of carbon dioxide in the carbon cycle. This compound is required for photosynthesis, is one of the products of respiration, and is a component of the sedimentary process in the form of carbonates, shells of organisms, or carbon in organic detritus. The last may provide energy for the benthos long after primary production has ceased, be buried to enter later into the longer geochemical cycle, or perhaps even be transported far from its place of origin by ocean currents to play a significant role in the global carbon cycle. Fig. 22-1 is a simplified carbon

Figure 22-1. Simplified flow of carbon in the ecosystem. The vertical shaded bar represents the division between the inorganic and organic part of the carbon dioxide system.

cycle, showing only the essential features. The cycle can be divided into several compartments: gas exchange through the sea surface as controlled by Henry's Law; the dissolved inorganic components as controlled by the chemical equilibrium of the carbon dioxide system, including the carbonate minerals; photosynthesis or primary production, the major mechanism for carbon dioxide transfer from the inorganic to the organic carbon compartment as controlled by physiological requirements of plants and the chemical and physical structure of the water column; respiration, which returns organic carbon consumed by marine biota to the inorganic system as CO_2 , as controlled by biological energy transfer processes; formation of detritus as derived from the excess of primary production over consumption, fecal pellets, and animal remains (much of this detritus settles to the bottom as the organic component of particulate matter and provides food for the benthic biomass or enters into other sedimentary processes); and, finally, the alkaline earth carbonate secretions and precipitates which may form long-term deposits of carbon dependent upon solubility equilibrium relationships for recycling to the inorganic pool.

Since until now I have had only two opportunities to make carbon dioxide measurements in this region—one in May 1976 and one in June and July of 1978—the results reported here are preliminary in nature. And yet these observations have revealed some facts and raised some questions that should be reported now. For 1980 and subsequent seasons, more definitive and extensive studies are planned which it is hoped will permit a more rigorous evaluation of the

carbon budget, particularly that of the inorganic system.

METHODS

Samples taken on the *Acona* cruise in 1976 were stored and frozen, and nutrient determinations were made by Technicon Autoanalyzer techniques at the Seward Station of the Institute of Marine Science. On the *T. G. Thompson* cruise analysis was made aboard ship by the five-channel Technicon Autoanalyzer using the methods of Patton and Whitledge (PROBES Progress Report 1978).

The techniques for $p\text{CO}_2$ measurement in surface water have been described earlier (Ibert and Hood 1963, Gordon et al. 1973, and Kelley et al. 1971). During these experiments a stream of seawater pumped from a bow intake system at a rate of 10-20 l/min. was split so that 5 l/min. passed through an equilibrator system to provide a gas sample in equilibrium with the sea water phase (Hood and Kelley 1976). The large volume of sea water minimized temperature excursions of the sample to less than +1 C as it passed through the ship to the laboratory. Three to four liters of this water were run through the multiphase equilibrator (Ibert and Hood 1963, Hood and Kelley 1976) to assure complete equilibration with the gas stream without distorting the CO_2 equilibrium of the water. The CO_2 concentration in the dry air and in the dry air after equilibration with sea water is expressed as a volume fraction (mixing ratio) in ppm by volume when the total pressure exerted by the air passing through the infrared analyzer is at one atmosphere. In an approximation sufficiently accurate for this investigation, the barometric pressure was assumed to be one atmosphere. The concentration of CO_2 in the air phase was measured by relating the signal of the output of the dispersed beam IR analyzer (Beckman model 310) to that obtained from two standard gases (one above and one below analysis values). The standard gases were referenced to those of Keeling of Scripps Institute of Oceanography, La Jolla, California, accurate to ± 0.2 ppm. The accuracy of the technique employed here, on the order of ± 5 ppm, could be improved easily by eliminating some technical problems of instrument stability and the difficulty of reading concentrations from strip-chart recorder records.

The carbon dioxide exchange rates of CO_2 invading the sea surface were determined by using a free exchange curvette technique similar to that of Sugiura et al. (1963) and further described in Hood and Kelley (1976). The transfer rates obtained by this technique are probably on the low side since the

direct effect of wind on the sea surface is influenced by the curvette cover. Since actual transfer is a molecular diffusion process through an air-water film and a water-water film, the rate is probably determined by the liquid boundary, according to the theory of Bohr (1899). The experimental design of the measurements made here attempts to avoid influencing water turbulence and thus the thickness of the water-water layer. If results obtained by this method are indeed low, how low remains to be determined (see Skirrow 1975 for further discussion).

Measurements of pH were made according to the method of Smith and Hood (1964) using a single probe calomel-glass electrode and a Coleman null-point pH meter. Tris-buffer made up in sea water and adjusted to the temperature of samples in a water bath kept at sea-surface temperature was used as a pH standard.

Total CO_2 measurements were made by infrared measurement of the quantity of CO_2 released from acidification of 3 ml of a sample of sea water. Both peak height and peak integration techniques were used to estimate the concentration as compared to a sodium carbonate standard. Bottled nitrogen was used as the carrier gas to sweep the generated CO_2 from the sample through the IR analyzer. A precision of ± 3 per cent, based on replicate analysis, was obtained with this technique.

RESULTS AND DISCUSSION

The salinity, temperature, and density profile of a typical station in the study area is shown in Fig. 22-2. The detailed data for this station, including oxygen, nutrients, and pH, are shown in Table 22-1. The $p\text{CO}_2$ pressure of the surface water (2 m) was $162 \mu\text{A}$ and the air was $332.6 \mu\text{A}$ to give a $\Delta p\text{CO}_2$ of -170.7 . The stability of the water column was apparently sufficient to support primary production in the euphotic zone (above 35 m) as indicated by the low nutrients, high pH, and low $p\text{CO}_2$, but this zone was underlain with unusual finestructure with water properties gradually changing to those expected below a well-defined pycnocline. Kinder and Schumacher (Chapter 4, this volume) have reviewed the hydrography of this region; they observed that the outer Bristol Bay transition zone (PROBES study area) contains three ocean fronts between which the horizontal salinity gradient is near zero. The outer front, centered over the shelf break, has a horizontal salinity gradient of about 9.4×10^{-3} g/kg/km, similar to that of the middle front, centered over the 100-m contour. The inner front is at about the 40-m contour and is typified by a homogeneous vertical water mass; waters inside the front are heavily influenced by coastal

Figure 22-2. Temperature, salinity profile of typical station (No. 28, Cruise No. 226, R/V *Acona*, May 1976) in PROBES study area.

processes. The fronts appear to act as flux boundaries across which only limited amounts of materials are exchanged. The fine structure in the deeper water between the middle and outer fronts consists of layers of waters of significantly different sigma-t values that originate from the other fronts and penetrate the inner area as fingers. The significance of this phenomenon in providing the euphotic zone with nutrients has recently been examined by Coachman and Walsh (1980). Its importance to the CO₂ budget is being examined during the 1980 field season.

As seen in the results of the surface pCO₂ distribution study for May 1976 (Fig. 22-3), in the region of the shelf north of 55° and east of the 200-m contour the water was depleted of CO₂ by more than 70 μA and in most cases was more than 170 μA lower than the air. In general, west of the 200-m contour and south of 55° the pCO₂ of the water was within ±20 μA of that of the air. While the surface water of the region just north of Unimak Pass had slightly higher pCO₂ values than the air, which may indicate upstream upwelling or mild vertical mixing in the pass, no values on the order of +200-300 μA, typical of waters welled up from below the discontinuity layer, were observed (Hood and Kelley 1976).

To compare the surface pCO₂ values with other parameters measured in the euphotic zone, average values obtained to a depth of 35 m in a cross section of the widest part of the shelf at 55°50'N were plotted (Fig. 22-4). The observations made by Coachman and Charnell (1977) found the middle front well developed at the 100-m isobath and the outer front at about the 150-m isobath. The data shown in Figs. 22-3 and 22-4 for pCO₂ and nutrients, particularly nitrate, from the 1976 cruise clearly reflect the outer front but show no clear indication of the inner front. A similar situation prevailed for 1978 (Fig. 22-5); a slight increase in surface pCO₂ was seen on the eastern side of the front at a depth of approximately 100 m. Perhaps if data had been collected further shoreward the influence of the inner front would have been more evident; however, it is apparent from these data that the productivity regime which influences nutrient and pCO₂ values occurs throughout the shelf domains.

TABLE 22-1

Depth m	Temperature °C	Salinity ‰	Sigma-t	Oxygen ml/l	PO ₄ μgA/l	NH ₃ μgA/l	NO ₃ μgA/l	SiO ₃ μgA/l	pH
5	2.69	32.148	25.67	9.87	0.40	0.3	0.0	15	8.44
10	2.62	32.153	25.68	9.24	0.44	0.3	0.0	17	8.40
15	2.60	32.157	25.69						
20	2.57	32.154	25.69	9.09	0.56	0.3	0.0	16	8.42
25	2.52	32.152	25.69						
30	2.47	32.152	25.69	9.08	0.44	0.3	1.7	16	8.37
35	2.42	32.120	25.67						
40	1.85	32.120	25.71						
45	2.06	32.363	25.89						
50	2.07	32.560	26.05	8.45	0.86	1.2	6.2	18	8.36
75	2.67	32.563	26.01	8.17	0.94	1.7	8.3	22	8.32
100	2.81	32.872	26.24	6.79	1.72	1.1	23.6	51	8.18

Figure 22-3. Distribution of $p\text{CO}_2$ in surface water in PROBES study area, May 1976.

Nitrate-chlorophyll- $p\text{CO}_2$ relationships

A plot comparing the average nitrate values of the top 35 m of the 1976 study area with $p\text{CO}_2$ is presented in Fig. 22-6. The circled dots in the top figure show zero nitrate values at the surface where $p\text{CO}_2$ was measured, but higher values at depth. The clustering of points of near-zero nitrate values at $\Delta p\text{CO}_2$ values of less than $-135 \mu\text{A}$ in the top figure, which represents 1976 data, and $-110 \mu\text{A}$ for the middle figure, which represents 1978 data, is of considerable physiological interest. These data indicate that the photosynthesizing plants are capable of removing all the nitrate from the water column relatively uninfluenced by the concentration of molecular carbon dioxide. It may be that other forms of CO_2 (i.e., HCO_3^-) are being utilized; if so, the stoichiometric effect on the $p\text{CO}_2$ is the same as for direct utilization (Fig. 22-1). The chlorophyll *a* content appears to bear little direct relation to $p\text{CO}_2$ content, probably because these two physiologically important components are functioning on different time scales. A high chlorophyll content is associated with a low $p\text{CO}_2$, but the recovery time for CO_2 is slow (because of limited atmospheric exchange capacity and depen-

dence on respiration of fixed carbon for recovery) and the photosynthesis which caused its decrease had ceased to occur long enough before the analysis was made that the plants and their chlorophyll *a* had disappeared. The utilization of ammonia, present in

Figure 22-4. Cross section $55^{\circ}50'N$ eastern Bering Sea shelf in June 1976. All parameters av. 0-35 m depth except $\Delta p\text{CO}_2$, which is 2 m.

Figure 22-5. Distribution of $p\text{CO}_2$ in surface water in five cross sections of PROBES study area in 1978.

the euphotic zone in concentrations of $0.1\text{--}1.0 \mu\text{gA/l}$, is associated with high chlorophyll a , but the biological processes which released the ammonia from organic matter would be expected also to release carbon dioxide to the water column in about the same proportion as it was fixed by photosynthesis.

The $p\text{CO}_2$ data, as well as all other data (see Hattori and Goering, Volume 2), suggest that nitrogen limits phytoplankton growth in this system and that the recovery of CO_2 concentrations is relatively slow compared to the dynamics of photosynthesis. Maintenance of the low $p\text{CO}_2$ values for the periods found in these studies must be related to a slow resupply of CO_2 to the euphotic zone, controlled by the exchange rate of CO_2 through the sea surface, vertical transport or respiration. To maintain low $p\text{CO}_2$, then, photosynthesis must be supported by nitrate, not ammonia, through resupply from below the discontinuity layer. Seasonally, of course, nitrate is resupplied from the mixing associated with thermal-haline convection.

Figure 22-6. Relation between nitrate-nitrogen concentrations and difference between water and air $p\text{CO}_2$ values. Top: data for 1976 R/V *Acona* cruise. Middle and bottom: data for 1978 Leg III R/V *T. G. Thompson*.

Total carbon dioxide

During the *Thompson* cruise in 1978, a preliminary investigation of the total carbon dioxide in the water column was undertaken. The vertical distribution at three stations in the PROBES area is shown in Table 22-2. Station 3016 was in the outer frontal area ($55^\circ 33.8'N$, $168^\circ 10.4'W$), station 3044 ($55^\circ 53.0'N$, $165^\circ 13.1'W$) in the central domain, and station 3086 ($56^\circ 56.8'N$, $166^\circ 54.0'W$) in the middle frontal area in a region of high chlorophyll a . All the values found were low compared to the concentration of 2.05 mM normally found in surface sea water during unproductive periods in this region of the world's ocean (Park et al. 1974). Station 3044 was in an area

TABLE 22-2
Total CO₂ at stations 3016, 3044, and 3086
mM CO₂/l

Depth	Sta. 3016	Sta. 3044	Sta. 3086
0	1.94	1.45	1.63
5	1.96	1.37	1.61
10	1.85	1.35	1.63
15	1.93	1.50	1.64
20	1.91	1.54	1.58
25	1.85	1.61	1.59
30	1.99	1.64	1.68
40	1.94	1.69	1.66
50	2.05	1.75	1.70
75	1.95	1.71	1.65
100	1.99	1.74	—

of low nutrients in which the primary production peak had already passed. The surface values found were significantly lower than at station 3086, in an active primary production area. At station 3016, in the outer frontal zone, a higher level of total carbon dioxide appears to be retained. Few conclusions can be drawn from these limited data regarding the distribution of total CO₂ on the eastern Bering Sea shelf, but it is clear that the quantity of total CO₂ varies with environmental conditions and ecosystem activity. It is also clear that the loss of approximately 450 g of inorganic carbon from the water column between the outer front, suggested to be at near pre-bloom concentrations of total carbon dioxide, and the middle domain, representing conditions after the spring bloom, must be accounted for. (See Kinder and Schumacher, Chapter 4, this volume, for location of these domains.)

A detailed study of the total CO₂ in the PROBES area has been undertaken during this 1980 field season (Hood and Codispoti, personal correspondence) to clarify the questions raised here concerning the loss of total carbon dioxide and relate its chemistry to the ecosystem processes of the shelf region.

Diurnal changes

On one occasion during the *Acona* cruise in 1976, a 24-hour station was occupied at 167°05'W, 55°8'N. Measurements of pCO₂ were made at 30-minute intervals except for two-to-three-hour periods when the equipment was being used for measuring the rate of exchange of CO₂ through the sea surface. It was expected that during the hours of darkness (between 2200 and 0300 local time at this latitude in May) the

large negative $\Delta p\text{CO}_2$ between water and air would be reduced by the continued flux of CO₂ from atmosphere to the water and by respiration of the indigenous organisms during a non-photosynthetic period. Extreme diurnal shifts in pCO₂ in the water were earlier observed by Park et al. (1958) over seagrass meadows in Texas bays and by Kelley (1975) over eelgrass meadows in Izembek Lagoon, Alaska.

The results of this study (Fig. 22-7) do not support the diurnal shift concept. The sudden increase in pCO₂ which occurred between 0430 and 0500 was probably caused by an influx of a different water mass at this moment, as evidenced by an accompanying drop in temperature, a slight increase in nitrate, and a slight change in the fine spectrum of the water column (Nishiyama 1976). Moreover, a diurnal change would probably have been gradual rather than sudden and would probably have coincided with the hours of darkness rather than coming well after dawn.

During the third leg of the *T. G. Thompson* cruise in May-June 1978 several stations were revisited after a time lapse of 9-13 days. The results of these observations are given in Table 22-3. The first four paired stations listed (6, 7, 8, 9 and 98, 97, 95, 84) were in the northeast section of the outer shelf domain and across the middle front into the central shelf domain (see Kinder and Schumacher, Chapter 4, this volume). At the time of the first visit these stations all had significant quantities of nitrate-nitrite in the euphotic zone; this was depleted during the period between the visits, and yet only moderate decreases in pCO₂ were observed at three paired stations while at the fourth (7 and 97) pCO₂ remained essentially the same. At stations 10, 11, 12, and later 83, 81, and 80, nearly zero nitrate-nitrite was observed on both visits and

Figure 22-7. Surface water pCO₂ and pH values, averaged to 35 m depth, at a 24-hour station in PROBES area in May 1976.

TABLE 22-3

A comparison of $\Delta p\text{CO}_2$ at the same geographic locations
after a time lapse of several days

Sta.	Location	Time	$\Delta p\text{CO}_2$	$\text{NO}_3^- + \text{NO}_2^-$ $\mu\text{gA/l}$	Sta.	Location	Time	$\Delta p\text{CO}_2$	$\text{NO}_3^- + \text{NO}_2^-$ $\mu\text{gA/l}$	Elapsed time
6	56°38'9N; 168°03'0W	0100 May 28	-110	20.3	98	56°38'1N; 168°05'0W	1100 June 8	-113	0.0	11.5 da.
7	56°50'9N; 167°51'9W	0330 May 28	-114	17.8	97	56°51'1N; 167°51'9W	0800 June 8	-110	0.0	11.3 da.
8	57°08'6N; 167°36'6W	0730 May 28	-76	5.8	95	57°10'0N; 167°33'0W	0430 June 7	-88	3.0	10.9 da.
9	56°56'0N; 166°54'0W	1200 May 28	-121	11.8	84	56°56'0N; 166°53'9W	0600 June 7	-140	0.0	9.8 da.
10	56°56'0N; 167°4'0W	1500 May 28	-142	0.1	83	56°44'0N; 167°3'9W	0300 June 7	-136	0.0	9.5 da.
11	56°31'6N; 167°16'9W	1830 May 28	-143	0.0	81	56°31'4N; 167°16'9W	2300 June 6	-131	0.0	9.2 da.
12	56°21'1N; 167°27'1W	2000 May 28	-144	0.1	80	56°21'0N; 167°27'8W	2000 June 6	-121	0.0	9.0 da.
16	55°33'8N; 168°10'4W	0731 May 29	-96	6.3	117	55°34'0N; 168°09'0W	1030 June 11	-100	3.6	13.2 da.
18	55°20'0N; 168°20'0W	1658 May 29	-88	3.8	118	55°39'9N; 168°13'1W	1130 June 11	-108	1.9	13.0 da.

after approximately nine days elapsed the $p\text{CO}_2$ did increase, although not dramatically. Stations 16 and 18 and later 117 and 118 show a decrease in nitrate-nitrite as well as $p\text{CO}_2$ after the 13-day time lapse.

In general the $p\text{CO}_2$ of the surface water tends to fall with the utilization of nitrate and nitrite (bloom conditions) and increase in waters with no nitrate and nitrite nutrients. It is clear, however, from these studies that a depressed $p\text{CO}_2$ in the water column requires considerable time to regain equilibrium with the atmosphere. The slow recovery is directly related to respiration and the exchange of CO_2 between the atmosphere and the ocean, since these are the primary sources of carbon dioxide to the ocean.

Carbon dioxide exchange

The environmental conditions found in late May 1976 and again in late May and early June 1978 produced very low $p\text{CO}_2$ values in the surface waters,

perhaps the lowest ever measured in the open ocean. From our general knowledge of the carbon cycle these low values could only be produced by an excess of photosynthesis over respiration, the rate of which is related to the flux of CO_2 into the sea surface. This rate is expressed by Henry's Law:

$$F = \alpha \Delta p\text{CO}_2 A \quad (1)$$

where F is flux of gas in moles/ m^2 /day, α is the exchange coefficient for CO_2 through the sea surface in moles/ m^2 /day, $\Delta p\text{CO}_2$ is the difference in atmospheres between the $p\text{CO}_2$ of the water and the overlying air, and A is area in square meters. The exchange coefficient, α , is very difficult to measure in the field because of the changing sea-surface conditions and the requirement that the complex surface, active in exchange, not be disturbed by the experimental conditions. Direct measurements of α have been attempted by Sugiura et al. (1963), Park and Hood (1963), and Hood and Kelley (1976). The

basic technique used in these three papers, reviewed by Skirrow (1975), employs a canopy-type capsule placed on the sea surface, allowing free movement of the water phase while the capsuled air phase is sampled for incremental concentration changes which are related to the exchange coefficient. The fact that the method limits the immediate wind effect on the sea surface probably lowers the exchange rate and thereby renders minimum values for the exchange rate α . Broecker and Peng (1974) have estimated the exchange rates for the world ocean to be $792 \text{ g/m}^2/\text{d/atm}$ (1700 m/yr piston velocity) based on radioactive carbon distribution data.

A summary of the exchange data reported in the literature is given in Table 22-4, along with an estimate of the flux in grams of carbon exchanged per day under Bering Sea conditions. The time required for the Bering Sea shelf waters to reach equilibrium with the atmosphere can be estimated if the exchange rates are known. Chosen for consideration are the conditions represented by the time series stations 10, 11, and 12 (occupied 8 May 1978) and 83, 81, and 80 (occupied 6 June 1978) (see Table 22-3). These six stations represent three pairs at the same geographical locations occupied about nine days apart. The water conditions, including nutrients at near-zero concentrations, were essentially the same on both visits, indicating post-bloom status. The initial pCO_2 of the surface water averaged $187 \mu\text{A}$ or $143 \mu\text{A}$ below that of the atmosphere. The total CO_2 de-

ficiency in the water column is assumed to be the same as given in Table 22-2 or about 37.5 M of CO_2 less than in pre-bloom conditions. At an estimated invasion rate of CO_2 based on the $792 \text{ g/m}^2/\text{d/atm}$ exchange rate and a pCO_2 of $143 \mu\text{A}$, the flux rate is $1.1 \text{ g/m}^2/\text{d}$. The deficit in total CO_2 for the 100 m of the water column from Table 22-2 is about 490 g. Since equilibrium will be approached in a logarithmic manner characteristic of the changing partial pressure, the time to reach one-half equilibrium would be 490 g deficiency divided by 1.1 g flux/d multiplied by 0.692 or 308 days, assuming that water chemistry remains the same.

During the nine days between the observations at the time series stations, 10 g of CO_2 (based on seven-year residence time) should have invaded the sea surface, representing a 2-percent recovery of the deficit in total CO_2 . The change in measured pCO_2 was from an average of $-143 \mu\text{A}$ at the beginning to $-130 \mu\text{A}$ after nine days, or a change of 8 percent, for part of which respiration must be responsible. Even considering the many assumptions and errors in experimental measurements, these data nevertheless indicate that short-term changes in the CO_2 system should not be expected since the rate of supply from the atmosphere is slow in comparison to the deficit in CO_2 caused by the spring bloom and low rates of respiration in the water column. The absence of a short-term shift in pCO_2 which should be brought about by respiration must be the result of relatively small animal populations and little microbial activity under these conditions, in comparison to the deficit in total carbon dioxide concentrations created by the earlier heavy phytoplankton bloom.

Ultimately animals in the water column or in the benthos will consume most of the fixed organic carbon and return the inorganic carbon components to the system, thereby greatly speeding up the rate at which equilibrium with the atmosphere is reached. Only fixed carbon lost to the sediments, stored in biomass, or transported from the area would be replaced by CO_2 from the air. Carbonate precipitation would also influence the balance but its importance has not been evaluated here since alkalinity measurements, useful for estimation of changes in carbonate ion concentration, were not available in these studies.

Influence of birds and mammals

The influence of the large numbers of birds and mammals which consume fixed carbon from the sea and release much of it directly to the air through respiration, or to land and sediments through defecation, can at the present time only be estimated, since population data for the PROBES area are not yet

TABLE 22-4

Exchange rate coefficients and flux of CO_2
through the sea surface

Reference	α		Flux* $\text{g/m}^2/\text{day}$
	Media	($\text{g/m}^2/\text{day/atm}$)	
Bohr 1899	Lab	3740	0.60
Suguira et al. 1963	Lab	4610	0.74
Hoover and Berkshire 1969	Lab	2160	0.35
Miyake and Hamanda 1960	Lab	290	0.05
Guyer and Tobler 1934	Lab	94320	15.0
Seven-year atmospheric residence time	Fld	7920	1.3
Suguira et al 1963 (1)	Fld	28220	4.5
Suguira et al. 1963 (2)	Fld	13100	2.1
Station 3086, June 1978*	Fld	11000	1.8
Station 3118, June 1978*	Fld	8500	1.4
Station 3114, June 1978*	Fld	3450	0.6
Station 3114, June 1978*	Fld	2020	0.32

*Used a pCO_2 pressure difference between water and air of $160 \mu\text{A}$, the average for the cruise period of 1978.

available. It has, however, been estimated that 450,000 tons of marine mammals which gain 1 million tons of body weight per year inhabit the eastern Bering Sea shelf (PROBES 1973). If mammals are 10 percent efficient, then 10 million tons of food is consumed to achieve this gain in body weight. If food items are 30 percent solid matter and 50 percent of that is carbon then 15 percent of the food items or 1.5 million tons (1.5×10^{12} g) of carbon or 5.5 million tons of carbon dioxide are respired by marine mammals annually. If one-half of this is transported directly to air or land, for each m^2 of the shelf ($1.22 \times 10^{12} \text{ m}^2$) 4.5 g of CO_2 is transferred in this way to be replaced by atmospheric carbon dioxide through sea-surface exchange.

Birds effect carbon dioxide transfer in a similar way, but are estimated to have only 10 percent of the impact of mammals. Together these two groups may transfer 5.0 g $\text{CO}_2/\text{m}^2/\text{yr}$. By comparison, the commercial fish catch of about 2 million tons per year on the Bering Sea shelf removes only about 1 g $\text{CO}_2/\text{m}^2/\text{yr}$.

The carbon budget in estimating the flow of fixed energy

The gross primary production in a system may be expressed as

$$Pg = F + (\Sigma \text{CO}_{2(i)} - \Sigma \text{CO}_{2(t)}) + R \quad (2)$$

where Pg is gross productivity; F is $\alpha \Delta p\text{CO}_2$ or flux of CO_2 through the sea surface, as given in expression (1); ΣCO_2 is total inorganic carbon in the water column initially (i) and at time of computation (t) (ΣCO_2 in both cases, here and in following computations, must be corrected for carbonate precipitation, which is based on changes in alkalinity between initial and final time); and R is respiration.

The net productivity by one definition $P_{n(1)}$ is that production which occurs in excess of what is required to maintain the photosynthetic apparatus, or:

$$P_{n(i)} = F + (\Sigma \text{CO}_{2(t)} - R_p) \quad (3)$$

where all terms of the expression are the same as in (2) except R_p , which represents respiration by plants and associated organisms. This value is probably not obtainable in the field, but must be measured by laboratory bottle experiments. This net productivity is that normally measured by the conventional bottle techniques used in primary productivity measurements.

A second net production is represented by:

$$P_{n(2)} = F + (\Sigma \text{CO}_{2(i)} - \Sigma \text{CO}_{2(t)}) \quad (4)$$

This is that production which occurs in excess of respiration of the total biomass of the water column at

time t . The CO_2 component which is derived from respiration is effective in increasing total CO_2 and increasing $p\text{CO}_2$ of the water, thereby reducing the flux.

To estimate the energy flow to respiration, data are needed for different times during the growing season, in which case:

$$R_B = (Ft_2 - Ft_1) + (\Sigma \text{CO}_{2t_2} - \Sigma \text{CO}_{2t_1}). \quad (5)$$

Such data were not available from the 1976 and 1978 field season, but will be collected in future work. Some indication of respiration rate can be gained, however, from the exchange data for the time series at stations 10, 11, and 12 and 83, 81, and 80 (Table 22-3). During this period 10 g of CO_2 was exchanged through the sea surface based on average exchange rates for the world ocean and the $p\text{CO}_2$ of the water column increased by 8 percent. Since this percent increase is the same as that for total CO_2 (Broecker 1974), the deficit of CO_2 found at station 3044 of 1620 g/ m^2 would have changed by 130 g CO_2/m^2 (35 g C/ m^2). This amount plus the 10 g of carbon exchanged gives an estimate for respiration for the nine days involved of 5.0 gC/ m^2/day . This value appears high and should not be used as an estimate of respiration for budgetary studies until more data are available and the method is refined by direct measurements of both exchange rates and total CO_2 at the times of interest. It is given here only to indicate the potential of this technique in helping to understand energy flow in the ecosystem.

Another net productivity $P_{n(3)}$ that is more easily estimated and that is also valuable for energy flow considerations is the productivity or fixed carbon that remains in the system after grazing by animals in the water column. This might be considered stored energy to be used later by all higher trophic levels of the system. It is represented by the expression:

$$P_{n(3)} = \Sigma \text{CO}_{2(i)} - \Sigma \text{CO}_{2(t)} \quad (6)$$

and can be determined for any time when the total carbon dioxide content of the water column is known. For example, using data from Table 22-2 for station 3016 as $\Sigma \text{CO}_{2(i)}$ and station 3044 as $\Sigma \text{CO}_{2(t)}$, the difference in integrated values for these two stations ($P_{n(2)}$) is 37 moles; about 450 g of fixed carbon/ m^2 is held in the water column for future utilization. To provide reliable data for this parameter, total carbon dioxide measurements before and after the spring bloom at enough stations to allow estimates of net productivity for the whole PROBES area will be needed.

The partitioning of $P_{n(3)}$ into compartments of energy for use by the various components of the biological community will require that the distribu-

tion of particulate carbon over the study area be determined. These numbers coupled with inorganic carbon studies will provide data on the amount of fixed carbon reaching the benthos as well as how much might be horizontally transported from the area.

SUMMARY AND CONCLUSIONS

The data reported here are of a preliminary nature and, except for $p\text{CO}_2$ data in the surface water, do not permit us to draw reliable conclusions. It is well established that a large area of the eastern Bering Sea shelf annually becomes extremely depleted in gaseous CO_2 with respect to the overlying air mass. This depression is associated with primary production and is accompanied by a decrease in nutrients and total carbon dioxide components. Once the depression is established, recovery to equilibrium is slow. The deficit in total CO_2 caused by the spring bloom (on the order of 450 gC/m^2) can only be replaced by atmospheric exchange (on the order of $1.5\text{--}4.0 \text{ gC/m}^2/\text{d}$) and respiration (on the order of $2\text{--}5 \text{ gC/m}^2/\text{d}$) or between 1 and 2 percent per day.

On the basis of the information obtained thus far, it is clear that understanding the carbon budget can be valuable in determining energy flow in the course of PROBES ecosystem studies and will lead to a much better understanding of the amount of carbon fixed in the system and where this carbon is going within the system.

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Organic Matter in the Bering Sea and Adjacent Areas

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ABSTRACT

Particulate matter collected in the Bering Sea and its adjacent areas during the cruises of R/V *Hakuho Maru* in the summers of 1975 and 1978 was analyzed for organic carbon and nitrogen and chlorophyll *a*. The concentrations of particulate organic carbon and nitrogen were measured with the ranges of 16-420 $\mu\text{gC/l}$ and 1-85 $\mu\text{gN/l}$, 19-185 $\mu\text{gC/l}$ and 1-25 $\mu\text{gN/l}$, 46-1,040 $\mu\text{gC/l}$ and 6-160 $\mu\text{gN/l}$, and 82-380 $\mu\text{gC/l}$ and 16-70 $\mu\text{gN/l}$ in the Oyashio, the deep Bering Sea, the continental shelf, and the Chukchi Sea areas respectively. In view of the concentration of particulate carbon and nitrogen, the most productive areas in the Bering Sea are the Oyashio area and the transition areas of the continental slope to the continental shelf, where inorganic nutrient supply systems are well operated by the East Kamchatka Current and upwelling respectively. The ratios of particulate organic carbon to chlorophyll *a* were in agreement with these conclusions.

The particulate matter was also analyzed to determine the amino acid, monosaccharide, and fatty acid composition. Organic compounds such as essential amino acids, mannose and glucose, and saturated fatty acids are suggested as valuable tracers for analysis of vertical transportation of organic matter in the marine environment. Polyunsaturated fatty acids were implicated as useful diagnostic tools in determining the biological activity of phytoplankton in oceanic areas. Vertical variability of the organic composition is briefly discussed with relation to the biological degradation of particulate organic matter.

INTRODUCTION

The Bering Sea is one of the largest confined seas of the world ocean, bounded on the west and east by the land masses of Siberia and Alaska respectively and on the south by the Alaska Peninsula and Aleutian Islands arc (Shore 1966). It has been reported by several workers (Zenkevitch 1963, Hood and Kelley 1974) that the Bering Sea and its adjacent northern North Pacific Ocean are areas where some of the highest values of both primary and secondary biological productivity have been observed in the world oceans. This is thought to be due to the abundant

supply of inorganic nutrients provided by the influx of nutrient-rich water masses from the North Pacific Ocean into the Bering Sea and by extensive vertical mixing during winter.

Advancement of the hydrography of the Bering Sea has been made by extensive studies conducted by several workers over the last decade (Favorite 1967, 1974; Ohtani et al. 1972). In the Bering Sea the warm eastern subarctic Pacific water is transformed into cold western subarctic water (Ohtani 1970). Seventy-five percent of the water entering the Bering Sea originates from the Alaskan Stream (eastern subarctic Pacific water) and the rest from water of the subarctic gyre (Favorite 1974). After entering the Bering Sea, water is involved in a large counterclockwise gyral movement, which causes upwelling of deep water as observed in the temperature profile of the deep Bering Sea basin.

Another remarkable characteristic of the vertical structure of water throughout the Bering Sea is the dichothermal water which occurs in a well-defined intermediate cold layer below the surface (Uda 1955, Takenouti and Ohtani 1974). The surface water is cooled in winter and a dichothermal layer of uniform salinity and low temperature is formed by thermal convection. The upwelling of more marine water associated with the counterclockwise eddies increases the salinity and temperature gradients underneath the surface layer. During the summer the temperature near the surface increases, but the warming does not reach to the bottom of the surface layer where the cold water remains as a dichothermal layer.

Apart from the above studies, little is known concerning hydrographic features of the Bering Sea

which influence the distribution of organic matter. Only limited data are available on the distribution of dissolved and particulate organic matter in the Bering Sea. Loder (1971) and Hood and Reeburgh (1974) report some data for a region north of Unimak and Unalaska Islands in the eastern Aleutian arc and Nakajima (1969) gives data for a region of the eastern deep Bering Sea. The extent of horizontal and vertical variability in the concentration of dissolved and particulate organic matter is not well established.

The aims of this chapter are: first, to show the distribution profiles of dissolved and particulate organic carbon and nitrogen in the Bering Sea and its adjacent areas, including the Oyashio, the deep Bering Sea, the eastern continental shelf, and the Chukchi Sea, in order to find the factors which control the regional variability of these organic elements; and, second, to describe the biochemical constituents of particulate matter to gain a better understanding of the biological and chemical aspects of the dynamic processes which control the distribution of organic material in these oceanic areas.

MATERIALS AND METHODS

Five to ten liters of sea water samples were filtered through a glass fiber filter (Whatman GF/C) with 47 mm diameter, precombusted at 450 C for four hours. The particulate matter collected on the filter was kept frozen at -20 C until analysis. Before analysis, the filters were allowed to stand in a chamber filled with HCl vapor for one hour to remove carbonate-carbon, after which they were dried in an oven for two hours at 80 C. Particulate organic carbon and nitrogen were determined by CHN-Corder, Yanaco Model MTS-2 carbon-nitrogen analyzer.

Menzel (1966) proposed that dissolved organic matter was adsorbed on a glass fiber filter during the filtration processes. Thus, an estimate of the organic carbon and nitrogen adsorbed was conducted to correct the concentration of the particulate organic carbon and nitrogen. A sea water sample collected in the Mikawa Bay near Nagoya was filtered through glass fiber filters in various volumes separately. Each of the filters was analyzed for total organic carbon and nitrogen. The results obtained indicated that 38 μg and 11 μg of dissolved organic carbon and nitrogen respectively were adsorbed on each of the glass fiber filters during the filtration.

The concentration of chlorophyll *a* was determined by the fluorometric method described by Yanagi and Handa (1970).

The fluorometric method (Udenfriend et al. 1972)

was used for the determination of particulate amino acid after hydrolysis of combined amino acid with HCl (6N) at 110 C for 24 hours. This method was standardized on the composite sample of amino acids consisting of a bulk of *Chlorella vulgaris* (Fowden 1954). Content of carbon and nitrogen in amino acid was calculated by multiplying the amino acid value obtained by 0.475 and 0.139 respectively.

The amino acid composition of the particulate samples collected at stations 11 and 33 was determined by the amino acid analyzer, Hitachi Model KLA-5. Particulate matter collected from 40-50 l of sea water was collected on the glass fiber filter by filtration. The glass fiber filter was treated with 10 ml of HCl (6N) at 110 C for 24 hours to hydrolyze combined amino acids. The hydrolysate was evaporated to dryness at 40 C with a rotary evaporator. The residue was dissolved in 0.05N HCl, and applied to the chromatographic column. Acidic and neutral amino acids were separated by the column (9 mm in outside diameter, 250 mm in length) packed with Hitachi custom resin #2618 and eluted with sodium citrate buffer (0.2N), pH 3.25 and then pH 4.25. Basic amino acids were eluted from the column (9 mm in outside diameter, 100 mm in length), packed with Hitachi custom resin #2618, and eluted with sodium citrate buffer pH 5.28.

For the determination of fatty acid composition, 400-450 l of sea water were collected from various depths from the surface through 2,000 m and filtered through a precombusted glass fiber filter (Whatman GF/C) (285 $\times$ 420 mm).

The glass fiber filter with particulate matter was cut into small pieces, which were transferred into an Erlenmeyer flask. To the flask was added chloroform-methanol (2:1 v/v, 20 ml) and then a certain amount of methyl heneicosanoate ($\text{C}_{21:0}$) (usually 10 μg). The mixture was stirred vigorously under nitrogen at room temperature overnight. After removal of solid residues by filtration through a glass fiber filter, the filtrate was reduced to a small volume under reduced pressure by the rotary evaporator at 40 C. Distilled water (20 ml) was added and lipid materials were extracted with petroleum ether (45-60 C b.p., 20 ml) four times. The combined extracts were dried over anhydrous sodium sulfate and evaporated by the rotary evaporator and then dried *in vacuo*. To the dried material was added benzene (1 ml) and 0.5N KOH in absolute methanol (4 ml). The solution was refluxed under nitrogen at 100 C for two minutes to saponify lipid materials. Free fatty acids generated were methylated with 14 percent BF_3 in absolute methanol (5 ml) at 100 C for two minutes (Metcalf et al. 1966). After cooling, distilled water

(10 ml) and then petroleum ether (10 ml) were added to the reaction mixture successively and methyl esters of the fatty acids were extracted with petroleum ether four times. The combined extracts were washed with distilled water three times and dried over anhydrous sodium sulfate. The fatty acid methyl esters were cleaned up by thin-layer chromatography (TLC) using silica Gel G (Merck, type 60) and petroleum ether/ethyl ether/acetic acid (95: 5:1, v/v) as a solvent system for the development of the TLC plate. After extraction of the methyl esters from the silicic acid with chloroform-methanol (9:1 v/v, 10 ml), the extract was dried over anhydrous sodium sulfate and evaporated to dryness under reduced pressure.

The fatty acids methyl esters were dissolved in ethyl ether, and a few microliters of the solution were injected into the gas liquid chromatograph (Hitachi, Model K53) equipped with hydrogen flame ionization detector under the following conditions: glass column (2 m in length, 3 mm in inner diameter) packed with 1 percent OV-1 on Chromosorb W (80/100 mesh); carrier gas, nitrogen (flow rate of 30 ml/min.); oven temperature, 120-310 C programmed at 5 C/min.; injection port temperature, 300 C.

To the residue after extraction of lipid materials from particulate samples collected from various depths of the Bering Sea (as stated before) was added sulfuric acid (72 percent v/v, 5 ml). The glass fiber filter soaked with sulfuric acid was allowed to stand at room temperature for three hours and then hydrolyzed at 105 C for 20 hours after dilution to 1N sulfuric acid by the addition of distilled water. An aliquot of the hydrolysate was used for determining total carbohydrate by the phenol sulfuric acid method (Handa 1966). A certain amount of inositol was added to the hydrolysate as an internal standard for quantitative determination of carbohydrate and then the hydrolysate was neutralized with barium hydroxide. The barium sulfate formed was removed by centrifugation. The precipitate was washed with distilled water four times, and the combined supernatants were treated with sodium borohydride (ca. 0.05 g) to reduce sugars to corresponding sugar alcohols (Abdel-Akher et al. 1951). After 16 hours the excess sodium borohydride was decomposed by the addition of acetic acid until gas evolution ceased. The solution was evaporated to dryness. To the residue was added 10 ml of methanol and then the methanol was removed by evaporation under reduced pressure. This procedure was repeated four times to remove boric acid as a complex of boric acid with methanol. The residue was dehydrated completely *in vacuo* and then sugar alcohols were acetylated with acetic anhydride (15 ml) in the presence of

anhydrous sodium acetate (ca. 0.1 g) at 100 C for two hours. The reaction mixture was poured into ice water and allowed to remain at room temperature overnight with vigorous stirring. The sugar acetate was extracted four times with chloroform. The combined extracts were washed with distilled water, dried over anhydrous sodium sulfate and evaporated to dryness by the rotary evaporator. The residue was dissolved in a small volume of methylene chloride. A portion of the methylene chloride was injected into a gas liquid chromatograph (Yanaco Model G8) equipped with hydrogen flame ionization detector under the following conditions: glass column (1.5 m in length, 3 mm inner diameter) packed with 3 percent ECNSS-M on chromosorb W (100-120 mesh); carrier gas, nitrogen (flow rate of 30 ml/min.); oven temperature, 120-210 C programmed at 2 C/min.; injection port temperature, 300 C.

RESULTS

Distribution profiles of particulate organic carbon and nitrogen in the Bering Sea and its adjacent areas

Sea water samples were collected from oceanic areas indicated in Fig. 23-1 during the cruise of the R/V *Hakuho Maru* of the Ocean Research Institute, University of Tokyo, from 21 June to 18 August 1975.

The Oyashio area

Hydrographic stations were occupied at five locations off the Kuril Islands extending into the western basin of the Bering Sea. The concentration of particulate organic carbon in the surface water was measured in the range of 70 to 422 $\mu\text{gC/l}$, with 169 $\mu\text{gC/l}$ as an average value. High values of more than 200 $\mu\text{gC/l}$ of POC were found in the surface water at stations 4, 5, and 7, while low values of POC were measured in the surface water at station 6 (Fig. 23-2). Marked differences in the concentrations of POC among the stations were observed in the intermediate waters, where high values of POC were found at stations 4 and 7, but low values at station 6. These data clearly indicate that the concentration of POC in the surface water has an effect on that of the intermediate waters.

The distribution profile of PON is almost identical with that of POC (Fig. 23-3). The concentration of PON in the surface water was measured in the range of 17-85 $\mu\text{gN/l}$ as an average value. These values in the surface water tended to decrease with depth to one-half to one-third in the intermediate and deep waters, although regional variations in values were still evident in the deeper waters.

One of the most remarkable features of the

Figure 23-1. The cruise track of R/V *Hakuho Maru* in summer of 1975. Numbers represent stations occupied.

Oyashio/east Kamchatka current areas is the cold and less saline dichothermal layer at intermediate depth. The formation of the dichothermal water has been considered to occur first in the northeastern margin of the Bering Sea basin between Cape Olyutorsky and the Bay of Anadyr (Uda 1955, Ohtani et al. 1972). Strong clockwise eddies and currents along the coast and the continental shelf augment the sinking of the surface water to depth, resulting in the formation of a homogeneous water layer down to 500 m with very small temperature and salinity gradients. Precipitation and radiation in the summer cause the temperature of the surface waters to rise, leading to the formation of a cold water layer at intermediate depth in the Bering Sea and adjacent areas.

This dichothermal water flows from Anadyr Bay southwestward along the Kamchatka Peninsula as the east Kamchatka current and then along the Kuril Islands. Here it splits and part flows eastward as the subarctic current while the rest continues to flow southwestward as the Oyashio. A well-developed dichothermal layer with low temperature (-0.6 – 1.5 °C) and low salinity (32.9 – 33.8 ‰) was found to

occur in the intermediate layer at 50–311 m depth at stations 4 and 7, while higher values of temperature (2.0 – 3.7 °C) and salinity (33.1 – 33.9 ‰) were found at station 6. Temperature, salinity, and depth of the dichothermal water have been defined to be 1.0 – 2.0 °C, 33.7 – 33.8 ‰ (salinity at the bottom of the halocline) at 100–300 m depth for the east Kamchatka current and 1.0 – 2.0 °C, 33.8 ‰ at 100–140 m depth for the Pacific subarctic water (Ohtani et al. 1972). These results clearly indicate that stations 4 and 7 are in the east Kamchatka current, while station 6 is in Pacific subarctic water.

The vertical profiles of particulate organic carbon and nitrogen, chlorophyll *a* and some oceanographic elements such as temperature, salinity, and density are shown in Fig. 23-4 and 23-5. The concentration maxima of chlorophyll *a* were found at 20 and 30 m at stations 4 and 6 respectively. These depths coincide with the maximum concentrations of particulate organic carbon and nitrogen and with the minimum of POC/chl. *a* and C/N ratios at these stations. These data suggest that phytoplankton are actively growing at these depths. Values for POC/chl. *a* of 200 and C/N of >6 for the particulate matter

Figure 23-2. Distribution profile of particulate organic carbon in the Oyashio area.

Figure 23-3. Distribution profile of particulate organic nitrogen in the Oyashio area.

found in surface water layers at these two stations were observed. These data strongly suggest that the particulate organic carbon at these two stations consists largely of phytoplanktonic organic matter.

In the subsurface through deep waters, C/N values of particulate matter tended to increase with depth at station 6, while no significant change in the values with depth was found at station 4. These data indicate that the degradation of particulate organic matter by biological agents in the deep water is more active at station 6 than at station 4.

A characteristic feature of the waters in this area was found in the distribution profile of $\text{NH}_4^+\text{-N}$ (Fig. 23-4 and 23-5). The concentration maximum of

$\text{NH}_4^+\text{-N}$ was found to be below the euphotic zone at all the stations in this area. Higher values were found at stations 4 and 7, in the cold water masses, than at station 6, in the warm water masses. Total inorganic nitrogen consisting of $\text{NO}_3^-\text{-N}$, $\text{NO}_2^-\text{-N}$, and $\text{NH}_4^+\text{-N}$ was measured within the range of 20-30 $\mu\text{gN/l}$ in the waters above 100 m depth at both stations 4 and 7, but no significant difference in the concentration of total inorganic nitrogen between these stations was observed. Thus, the increment of $\text{NH}_4^+\text{-N}$ in the cold water masses might be due to the retardation of the nitrification process required to oxidize $\text{NH}_4^+\text{-N}$ to $\text{NO}_3^-\text{-N}$.

The deep Bering Sea area

The distribution of water temperature was observed at five stations in the deep Bering Sea. The surface layer above 20 m consisted of warm water at all stations, because of precipitation and radiation in summer as observed in the Oyashio area. The temperature then tended to decrease rapidly toward the deep to form an extremely sharp thermocline in the water layers between the surface and 50 m. Cold water with a minimum value of -0.57°C formed the intermediate layer of 50-200 m depth at stations 7 and 8. This indicated that the east Kamchatka current extended to station 8 at that time.

The dichothermal layer was much less developed at stations 9 and 10, while intermediate warm water was observed at depths between 250 and 500 m in the area centered at station 9. The 3.5°C isotherm showed a domelike upward curvature. Kitano 1970 reported that the warm water mass was introduced by the infiltration of the Alaska Stream into the Bering Sea across the Aleutian Island chain at Kamchatka Pass and west of the Attu Islands. We found intermediate water of temperature higher than 3.8°C in the area of $55\text{-}58^\circ\text{N}$ and $173\text{-}176^\circ\text{E}$, where we also found the intermediate warm water mass. These data indicate that warm water derived from the Alaskan Stream occurs consistently in the intermediate layer of the Bering Sea.

The distribution profile of salinity in the deep Bering Sea indicates that the surface and subsurface layers of stations 7 and 8 consist of waters characteristic of the east Kamchatka current and those of station 11 are affected by the intrusion of less saline water from the continental shelf area. The water between the surface and 250 m at stations 9 and 10 is not affected by the continental shelf water, but upwelling of the deep water is likely to occur in the area centered at station 9 because of the upward

Figure 23-4. Vertical profiles of particulate organic carbon and nitrogen, C/N, chlorophyll *a*, POC/chl. *a*, $\text{NH}_4\text{-N}$, temperature, salinity, and sigma- t at station 4.

curvature of the salinity profile from the surface to 500 m. These observations are in agreement with the observed temperature profile in this area.

The distribution profile of particulate organic carbon along 57°N in the deep Bering Sea is shown in Fig. 23-6 (see Fig. 23-1 for locations). The concentration of particulate organic carbon ranged from 80 to $240 \mu\text{gC/l}$ in the euphotic layer (0-50 m depth) (Otake et al. 1977). The values tended to decrease rapidly with depth below the euphotic zone at all stations except those on the continental shelf. The vertical distribution of particulate organic carbon in the intermediate water is rather complicated in this oceanic area. High concentrations of particulate organic carbon were measured in the intermediate cold waters at stations 7, 8, and 10, while low concentrations of particulate organic carbon were found in the intermediate warm waters centered at station 9.

The distribution profile of particulate organic nitrogen in the deep Bering Sea is shown in Fig. 23-7. More than $20 \mu\text{gN/l}$ of particulate organic nitrogen

was found in the euphotic layers at stations 7, 8, 12, 14, and 15. More than 10 was found at stations 9, 10, and 11. All these values tended to decrease to one-half at depths of 75-100 m. High values of particulate organic nitrogen were found in the cold waters of the intermediary depths at stations 7 and 8 as observed in the distribution profile of particulate organic carbon. These data indicate that low temperature retards the rate of decay of the particulate organic carbon materials by biological agents, resulting in high concentrations of particulate organic carbon in these waters.

Increased values of particulate organic nitrogen were found in the deep waters of stations 10 and 11. The transportation of particulate matter from the continental shelf along the continental slope may result in these high increments of particulate organic carbon and nitrogen in the deep waters.

A stepwise increase in the C/N value of the particulate matter was observed to occur in the water layers between 50 and 125 m and between 300 and 3,500 m (Fig. 23-8) at station 9. The C/N value was

Figure 23-5. Vertical profiles of particulate organic carbon and nitrogen, C/N, chlorophyll *a*, POC/chl. *a*, $\text{NH}_4\text{-N}$, salinity, and sigma-*t* at station 6.

found to be 6 in the euphotic layer, tending to increase rapidly to 10 with depth in the underlying waters, where an extremely rapid increase of the POC/chl. *a* value was also observed. The C/N value of the particulate matter tended to increase with depth to over 16 in the intermediate warm water at which depth the dissolved oxygen decreased to less than 0.5 ml/l. These data indicate that decay of particulate organic matter occurs in the intermediate as well as the deep waters of this region. No significance has yet been determined for the stepwise changes in POC or C/N values, but they are thought to be related to seasonal or yearly differences in production and distribution of organic matter.

The continental shelf area

The distribution profile of temperature in the transect along 57°N in 1975 over the continental shelf area indicates that a well-developed thermocline occurred in the water layers at approximately 30 m. Homogeneous distribution of temperature was observed in the layers above the thermocline throughout the stations, while a domelike cold water mass was found below the thermocline to the bottom in

the areas centered at station 14 where the core temperature was -0.5°C .

The formation and occurrence of the cold water mass has been described by several workers (Fleming 1955; Ohtani 1969; Kitano 1970a, b; Kinder and Schumacher, Chapter 4, this volume). The extremely cold water mass with core temperatures ranging from -1.0 to -1.7°C is formed in the area between the Gulf of Anadyr and St. Matthew Island south of St. Lawrence Island as a result of severe winter cooling. The cold water extends southeast while increasing its core temperature mainly by lateral mixing with surrounding warm water. The bottom water at station 14 may be the extreme southern front of the cold tongue since this feature was not found to the south at stations 19 or 20.

The vertical profile of salinity in the transect along 57°N indicates that water with relatively high salinity intrudes on the bottom of the continental shelf. However, the deep water at station 14 may not be affected by the intrusion of deep Bering Sea water. The waters of the eastern part of the continental shelf area were much affected by less saline coastal water and/or river water (Shore 1966, Roden 1967, Ohtani

Figure 23-6. Distribution profile of particulate organic carbon in the deep Bering Sea and the continental shelf areas.

Figure 23-7. Distribution profile of particulate organic nitrogen in the deep Bering Sea and the continental shelf areas.

1969). Water with salinity less than 31.5‰ occurred in the surface layers of stations 16, 17, and 18.

The distribution profiles of particulate organic carbon and nitrogen are shown in Fig. 23-9. Particulate organic carbon values higher than 100 µgC/l

were observed at all shelf stations with an increase eastward from station 12 to station 15 in the surface and subsurface layers, but not in the bottom waters. The distribution profile of particulate organic nitrogen in this area was almost identical to that of particulate organic carbon. C/N values of the particulate matter ranged from 6.2 to 8.5 at stations 12 through 16. Values of POC/chl. *a* ranging from 52 to 248 were observed in particulate matter at stations 12 through 16. These data indicate that the particulate organic matter consists of phytoplanktonic material to a large extent; however, no characteristic features of C/N and POC/chl. *a* were observed in core waters with low temperatures, as found centered at station 14.

Particulate organic carbon and nitrogen were found to be very abundant at station 17 and the Secchi disk depth was only 7 m (Otabe et al. 1977). The particulate matter of this station had C/N and POC/chl. *a* values of more than 15 and 700 respectively even in the euphotic zone. These data indicate that the standing stock of particulate matter was much affected by allochthonous materials at this station. Station 18 showed less effect of allochthonous materials on the standing stock of particulate matter than station 17.

One of the most characteristic features of the continental shelf areas was that the standing stock of inorganic nutrients was very low. SiO₂-Si and NO₃⁻-N were measured within the ranges of only 3-10 and 0.1-1.0 µg/l respectively and almost undetectable amounts of NO₂⁻-N and NH₄⁺-N were found in the water layers from the surface to 10 m at stations 12-18. High concentrations of particulate organic carbon and nitrogen were observed throughout this region. The inorganic nutrient concentrations tended to decrease from the oceanic areas (stations 11 and 19) to station 17, 10 miles south of Nunivak Island, with a steep gradient at the boundary between the continental shelf and the continental slope areas; conversely, concentrations of particulate organic carbon and nitrogen tended to increase toward the continental shelf.

Several authors (Koblentz-Mishke 1965, Karohji 1972, McRoy et al. 1972, Sanger 1972) have measured high productivity rates for the continental shelf area of the Bering Sea. It is clear that primary productivity of phytoplankton in the continental shelf area of the Bering Sea is nutrient dependent.

During the summer of 1978 a second cruise was made to the Bering Sea (Fig. 23-10). Distribution profiles of various materials along 173°-168° 30'W are shown in Fig. 23-11. Particulate matter was determined within the range of 0.25-4.38 mg/l in the

Figure 23-8. Vertical profiles of particulate organic carbon and nitrogen, C/N, chlorophyll *a*, POC/chl. *a*, temperature, salinity, and dissolved oxygen at station 9.

continental shelf area through the Chukchi Sea. Regional variability was evident. High values of particulate matter were detected in the surface and subsurface waters between the continental slope and St. Matthew Island areas, while low values were

observed between St. Matthew Island and the Bering Strait. Slightly higher values than in the Bering Strait were again observed in the Chukchi Sea. Almost identical distribution profiles were obtained for particulate organic carbon and chlorophyll *a*. More

Figure 23-9. Vertical profiles of particulate organic carbon (—●—●—), and nitrogen (—○—○—), C/N (—□—□—), chlorophyll *a* (—▲—▲—), POC/chl. *a* (—△—△—), temperature, and salinity at stations of the continental shelf area.

definite boundaries for the distribution of these materials were found to occur on the continental slope, St. Matthew Island, and Bering Strait areas. From these data, it can be seen that high productivity values due to phytoplankton occur in the area between the continental slope and St. Matthew Island and in the Chukchi area. Low productivity would be predicted in the area extending from north of St. Matthew Island to the Bering Strait.

No significant differences in the concentration of $\text{SiO}_2\text{-Si}$, $\text{NO}_3^-\text{-N}$, $\text{NO}_2^-\text{-N}$, $\text{NH}_4^+\text{-N}$, and $\text{PO}_4^{3-}\text{-P}$ between the areas were observed; thus the possibility that inorganic nutrient concentrations are the direct cause of regional variability in phytoplanktonic productivity is discounted. Vertical profiles of σ_t , however, suggest that upward transport of intermediate waters rich in inorganic nutrients occurred along the continental slope toward station 30. Distribution profiles of dissolved organic carbon (DOC) clearly show the occurrence of upwelling in the transition area between the continental slope and the continental shelf. From these data, it can be concluded that inorganic nutrients are being transported into the euphotic zone in the region of station 30, thus providing the high productivity observed. Poor phytoplanktonic productivity observed in the area between St. Matthew and St. Lawrence Islands is likely to be associated with the cold water mass in the bottom layer here. The water mass has a very steep temperature gradient (4 C/10 m) at 30-40 m, which limits mixing of the water and causes surface water nutrients during the spring bloom to be depleted with no later supply providing for continued primary productivity.

High productivity of phytoplankton was observed at station 21 in the Chukchi Sea, most likely as a result of the regional upwelling of sea water indicated by vertical distribution profiles of temperature and salinity.

A well-defined cold water mass centered at 100 m at station 6 was found to occur along the continental slope (Fig. 23-12 a and b); it was characterized by higher values of the ratio of particulate organic carbon to particulate matter (POC/PM) than were found in ambient waters. Salinity and temperature profiles indicate that the cold water mass is identical with the type D water mass described by Ohtani et al. (1972), formed by the mixing of Alaskan Stream waters flowing along the Aleutian Ridge and the continental slope. Considering that more than 30 percent of POC/PM was found in the Alaskan Stream waters off the Aleutian Islands arc (Tanoue and Handa 1979), it is conceivable that the particulate matter of the cold water mass at station 6 may be

Figure 23-10. Track chart of the KH-78-3 Cruise of the *Hakuho Maru*.

transported from the oceanic area south of the Aleutian Islands without significant modification in its chemical composition. Since there are no data on the chemical and biochemical nature of the organically rich particulate matter, it would be premature to discuss the processes of its formation.

DISCUSSION AND SUMMARY

Oceanic areas of the Bering Sea and its adjacent areas were divided into four regions on the basis of the hydrographic (Kitano 1970a and b, Takenouti and Ohtani 1974) and geographical (School et al. 1968) features. In this discussion similar areas were established and divided into the surface, intermediate, and deep water layers to provide for a more precise understanding of the characteristic features of the distribution of particulate organic matter. The results are shown in Table 23-1.

Regional variability was evident in particulate organic carbon and nitrogen determined to be in the ranges of 34-1,038 $\mu\text{gC/l}$ and 5-79 $\mu\text{gN/l}$ in the surface waters of the Bering Sea and its adjacent

areas. The average values of carbon and nitrogen tended to decrease in the following order: Chukchi Sea > the continental shelf > the Oyashio > the deep Bering Sea. The world's highest values of particulate organic carbon and nitrogen were found in the continental shelf and Oyashio areas. Such high values have never been detected in the open oceans except in the upwelling area of the South Pacific off Ecuador (Menzel 1967). The concentrations of particulate organic carbon and nitrogen in the Bering Sea and its adjacent areas were found to be two to five times higher than those obtained in the North Pacific Ocean (Holm-Hansen 1969, Gordon 1971, Handa et al. 1972), the North Atlantic Ocean (Chester and Stomer 1974, Banoub and Williams 1972, Gordon 1977), and the Indian Ocean (Menzel 1967, Chester and Stomer 1974). These high concentrations of particulate matter are considered to be among the most remarkable characteristics of the Bering Sea and its adjacent areas.

High concentrations of particulate organic carbon and nitrogen were not found in particulate matter collected from all of the oceanic areas of the Bering Sea, but were observed at stations 4, 5, and 7 in the

Figure 23-11. Distribution profiles of particulate matter, chlorophyll *a*, POC, and DOC along 173°-168°30'W.

Figure 23-12a. Temperature distribution in eastern Bering Sea in summer of 1978.

Figure 23-12b. Particulate organic carbon in percentage of total particulate matter in eastern Bering Sea.

Oyashio area and in the continental shelf area south of St. Matthew and Nunivak islands. The Oyashio area has a well-developed dichothermal in the intermediate layer of 50-250 m and the region south of St. Matthew is affected by upwelling of intermediate water along the continental slope from the deep Bering Sea. From these data, it can be seen that high values of particulate organic carbon and nitrogen were found only in areas where transport of inorganic nutrients to the euphotic zone occurred either because of vertical instability of the nutrient-rich intermediate water as observed in the Oyashio area or because of upwelling of the intermediate and deep

waters as observed in the continental shelf area.

To estimate the carbon content of living organisms in particulate matter, the relationship between chlorophyll *a* and organic carbon and nitrogen was examined. Linear relationships between these materials were found to occur and the regressions found for samples in the euphotic layers of the four areas are shown in Table 23-2. The independent terms of each of the equations are the concentration of the particulate organic carbon when chlorophyll *a* is equal to zero; this is the concentration of organic carbon and nitrogen in detritus or allochthonous materials. The ratios of the detrital organic carbon

TABLE 23-1

Summary of the ranges and average values (± 1 SD) of particulate organic carbon and nitrogen and C/N in the Bering and Chukchi Seas

Area	0-100 m Particulate organic			100-500 m Particulate organic			500 m Particulate organic		
	carbon $\mu\text{gC/l}$	nitrogen $\mu\text{gN/l}$	C/N	carbon $\mu\text{gC/l}$	nitrogen $\mu\text{gN/l}$	C/N	carbon $\mu\text{gC/l}$	nitrogen $\mu\text{gN/l}$	C/N
Oyashio area									
Range	70-422	10-85	3.3-8.4	28-142	2.2-21	5.4-26	16-145	1-21	5.5-20
N.S.*	8	28	28	43	42	42	41	37	37
Average	169 $\pm$ 81	29 $\pm$ 16	6.1 $\pm$ 1.0	62 $\pm$ 29	7.4 $\pm$ 3.6	8.8 $\pm$ 2.2	53 $\pm$ 27	5 $\pm$ 4	11 $\pm$ 2.8
Deep Bering Sea area									
Range	34-185	5.7-26	4.5-9.5	26-126	2.4-8.7	5.9-15	19-61	1.1-9.1	5.3-19
N.S.*	20	20	20	30	28	28	43	35	35
Average	91 $\pm$ 39	13 $\pm$ 6	7.1 $\pm$ 1.5	46 $\pm$ 21	4.5 $\pm$ 1.5	9.9 $\pm$ 1.9	33 $\pm$ 10	3.4 $\pm$ 1.9	11 $\pm$ 3.7
Continental Shelf area									
Range	50-1038	7.5-157	4.4-15	46-260	5.7-39	6.3-9.6	—	—	—
N.S.*	113	108	108	18	17	17	—	—	—
Average	204 $\pm$ 178	31 $\pm$ 25	6.4 $\pm$ 1.9	65 $\pm$ 44	7.4 $\pm$ 4.9	8.3 $\pm$ 4.9	—	—	—
Chukchi Sea area									
Range	82-376	16-72	48-59	—	—	—	—	—	—
N.S.*	15	15	15	—	—	—	—	—	—
Average	220 $\pm$ 94	42 $\pm$ 19	5.3 $\pm$ 0.28	—	—	—	—	—	—

*Number of samples.

TABLE 23-2

Relationship between chlorophyll *a* and particulate organic carbon in the euphotic layers of various ocean areas

Area	Number of Samples	Regression equation	r
Oyashio area	21	POC= 210 Chl. <i>a</i> + 64	0.714
	21	PON= 37 Chl. <i>a</i> + 8	0.814
Deep Bering Sea area	12	POC= 87 Chl. <i>a</i> + 48	0.671
	12	PON= 13 Chl. <i>a</i> + 10	0.866
Continental Shelf area	32	POC= 72 Chl. <i>a</i> + 94	0.458
	39	PON= 12 Chl. <i>a</i> + 11	0.616
Chukchi Sea area	12	POC= 60 Chl. <i>a</i> + 103	0.894
	12	PON= 12 Chl. <i>a</i> + 17	0.917

and nitrogen to the particulate organic carbon and nitrogen were calculated with the average values of 34.2 and 27.5 percent, 52.7 and 76.9 percent, 46.0 and 35.5 percent, and 46.8 and 40.5 percent for the Oyashio area, the deep Bering Sea area, the continental shelf area, and the Chukchi Sea area respectively. These data indicate that 50-70 and 25-70 percent of particulate organic carbon and nitrogen consist of

phytoplanktonic carbon and nitrogen in these oceanic areas. Holm-Hansen (1969) reported that the ratios of living organic carbon to particulate organic carbon were 62-73 percent and 55-73 percent on the basis of the biomass estimated by the measurements of ATP and chlorophyll *a* respectively. From these data, it can be seen that detrital organic carbon is abundant in the deep Bering Sea area, and phytoplanktonic organic carbon is more abundant in the particulate organic carbon of the Oyashio and Chukchi Sea areas.

Extremely high values of detrital organic carbon were found in the particulate matter from stations 17 and 18 off Nunivak Island on the continental shelf. Since less saline waters were observed at these stations, it can be assumed that drainage from land brings about the increase in detrital organic carbon of the particulate matter at these stations.

BIOCHEMICAL CONSTITUENTS OF PARTICULATE MATTER

Free and combined amino acids

Particulate samples collected at stations 11, 13, 14, and 33 in the deep Bering Sea, continental shelf, and northern North Pacific areas during *Hakuho Maru* cruise KH-75-4 were analyzed for free and combined amino acids. The particulate amino acids

were found to be within the ranges of 10.3-78.0, 104-156, and 10.4-96.4 $\mu\text{g/l}$ in the deep Bering Sea, continental shelf, and northern North Pacific areas respectively (Fig. 23-13). These values indicate that regional variations of particulate amino acids are identical with those of particulate organic carbon and nitrogen.

The concentration of particulate amino acids tended to decrease in a steep gradient with depth in the surface and subsurface layers at all stations. No significant vertical gradient was observed in the particulate amino acid of the deep waters at stations 11 and 33, but slightly higher values were found in the intermediate waters at station 33.

The ratios of particulate amino acid carbon (PAC) and nitrogen (PAN) to particulate organic carbon (POC) and nitrogen (PON) in the surface and subsurface layers (0-70 m) were found to have an almost identical range of values throughout the stations. Slightly lower values, however, were obtained at stations 13 and 14. It is assumed that this effect is

largely due to contamination by terrigenous materials since these stations are in Bristol Bay.

PAC/POC and PAN/PON at these stations were found with the ranges of 24.6-31.3 percent and 47.3-62.0 percent respectively. These values do not conflict with the ratios obtained in the high latitude areas (30-50°N) of the North Pacific Ocean, but much lower values were found in tropical and subtropical areas of the Pacific Ocean (Handa et al. 1972).

Amino acid composition of the particulate matter at stations 11 and 33 is shown in Table 23-3. Serine, glycine, and alanine were found to be the dominant components of particulate amino acids in the samples from all depths, whereas aspartic acid and glutamic acids dominated from the euphotic zone (50 m) down. Valine, leucine, and lysine were found in intermediate concentrations in all of the samples analyzed. The amino acid composition of the particulate matter collected from the surface was similar to that of marine phytoplankton, but the composition varied with depth. The molar ratios of alanine and arginine tended to decrease with depth, while those of serine, glutamic acid, and glycine increased.

Arginine, lysine, aspartic acid, glutamic acid, serine, glycine, and alanine have been reported as major amino acids in marine phytoplankton and in zooplankton and fecal pellets (Parsons et al. 1961, Siegel and Degens 1966, Starikova and Korshikova 1969, Daumas 1976). Arginine nearly disappeared from the water column in the deep waters of station 11, but was found at 2,000 m, the only depth sampled, at station 33. Rittenberg et al. (1963) and Degens et al. (1964) reported that arginine tended to decrease rapidly with depth in particulate matter and produce ornithine and urea (Clark et al. 1972, Degens 1970). We also observed 1.05-1.54 $\mu\text{g/l}$ of urea in particulate matter from the deep water at stations 11 and 33; samples were not analyzed for ornithine. Thus, these data may suggest that such a rapid decrease in arginine in particulate matter is due to bacterial use.

Relative abundances of the essential amino acids were much higher in the surface waters than in the deep waters, but significant amounts of the essential amino acids including arginine, lysine, histidine, isoleucine, and methionine were found in the particulate samples from deep water. In the particulate samples from the deep water of station 33, the relative abundance of the essential amino acids was found to be about the same as in the surface samples of particulate matter at station 11. These data strongly suggest that particulate amino acids in the deep water layers must be rapidly transported from

Figure 23-13. Vertical profiles of particulate amino acids at various stations in the deep Bering Sea, continental shelf, and northern North Pacific areas.

TABLE 23-3

Distribution of amino acids in particulate matter collected from various depths
(in residues per 1,000)

Station	Depth m	Basic			Acidic		Neutral								Sulfur Met	Aromatic Phe	Total mmol/l
		Arg	Lys	His	Asp	Glu	Thr	Ser	Pro	Gly	Ala	Val	Ile	Lue			
11	30	31	63	21	12	11	7	96	42	125	104	68	44	68	8	36	307
	50	32	86	29	92	75	61	170	35	150	95	58	35	23	6	23	278
	2,250	tr	60	17	106	129	53	132	40	182	83	59	40	69	3	30	121
	2,750	tr	69	20	109	132	59	125	43	164	86	63	39	63	3	26	122
33	2,000	34	73	26	87	97	49	171	24	191	73	46	101	12	34	24	234

the surface layers with only limited modification in transit. It appears likely that fecal pellets of zooplankton are an important carrier of the particulate amino acids from the surface to the deep water layers.

Carbohydrates

Particulate matter was collected for carbohydrate analysis (PCC) at several stations in the Bering Sea. Composite samples of this material were analyzed for carbohydrates by gas chromatographic techniques and found to be in the range of 15.3-32.1 $\mu\text{gC/l}$ or 13.1-27.2 percent of POC, in the continental shelf

area (Table 23-4). These values were about the same as the PCC/POC values (0.12-0.40) obtained in the Pacific Ocean from 50°N to 68°S along 170°W (Handa et al. 1972).

Regional variability in the concentration of particulate carbohydrate was evident. High values were found in the continental shelf area, whereas low values of PCC/POC were obtained in the deep Bering Sea. Parsons and Strickland (1962) reported that detrital samples from marine environments sometimes gave unexpectedly high concentrations of protein, while carbohydrate was only a minor component. McAllister et al. (1960) speculated that oceanic

TABLE 23-4

The monosaccharide composition of carbohydrate in particulate matter from various areas during the cruise of KH-75-4.

Monosaccharide	Deep Bering area		Continental Shelf area		
	Surface Layer (Stns. 9,10,11) 0-175 m	Deep Layer (Stns. 9,10,11) >175 m	Surface Layer (Stns. 12,16,17,19) 0-50 m	Bottom Layer (Stns. 12,19) 50-80 m	Bottom Resusp. (Stns. 19) 100-190 m
—mol. %—					
Phamnose	6.24	1.55	3.01	6.31	5.02
Fucose	19.1	7.03	3.33	7.81	5.37
Ribose	2.42	tr.	tr.	tr.	tr.
Arabinose	5.71	3.23	5.68	2.00	6.87
Xylose	3.90	2.55	2.80	5.80	11.0
Mannose	20.8	19.0	14.5	7.14	8.11
Galactose	5.48	1.25	5.39	65.2	10.6
Glucose	36.2	65.4	65.3		53.0
Total carbohydrate ($\mu\text{gC/l}$)	3.2	3.0	32.1	17.6	15.2
Carbohydrate carbon/ Org.C. (%)	3.91	8.42	13.5	27.7	13.1

detritus from the northeastern Pacific originated from fragments of zooplankton. These data suggest that the particulate matter of the deep Bering Sea may be largely composed of the bodies of protein-rich bacteria.

As shown in Table 23-4, glucose and mannose were found to be dominant species of monosaccharides after hydrolysis of the particulate carbohydrate with dilute sulfuric acid. Fucose was also found to be abundant, but only in the particulate samples from the surface and subsurface layers of the deep Bering Sea area. Comparable values of monosaccharides were obtained in the particulate samples from the continental shelf and the deep sea area although lower values of mannose were found in the shelf samples.

Handa and Yanagi (1969) reported that particulate carbohydrate of the northwestern Pacific Ocean was composed mainly of glucose and mannose, which accounted for 21.8-54.6 percent and 9.5-24.4 percent of the total carbohydrate. The authors found that fucose, also abundant, accounted for up to 12.8 percent of the particulate carbohydrate in those samples. On the basis of these data, there do not appear to be significant differences in the chemical nature of carbohydrates between the particulate samples from the Bering Sea and those from the northwestern north Pacific Ocean. According to detailed analyses of carbohydrates of marine diatoms conducted by Handa (1969) and Hang and Myklestad (1976), cell wall polysaccharides which are soluble in alkali consist mainly of mannose with smaller concentrations of fucose, glucose, galactose, rhamnose, xylose, and arabinose. Water-extractable polysaccharides, however, consist mainly of glucose, which accounts for 90 percent of this polysaccharide fraction. Arabinose, mannose, rhamnose, and ribose were detected as minor monosaccharide constituents. These data indicate that the high proportion of mannose to total carbohydrate in the particulate samples of the Bering Sea must be due to cell wall polysaccharides of diatoms which have been reported to be the main primary producers in the Bering Sea at the time when the particulate samples were collected (Ishimaru and Nemoto 1977).

Glucose, another main component of particulate carbohydrate, is also found in the intermediate and deep waters, but is unlikely to be derived from water-extractable polysaccharides of diatoms since it is quickly consumed by microbiological agents (Handa 1969). Handa and Yanagi (1969) reported that glucose still accounted for 30-41 percent of residual carbohydrate after removal of water-extractable carbohydrate in the particulate samples collected

from the surface and subsurface waters of the northwestern North Pacific Ocean. Thus these data suggest that a certain portion of the glucose found in the hydrolysate of particulate matter is derived from cell wall polysaccharides of the diatom cells.

Another source of glucose is terrigenous cellulose fibers which have been observed in the detritus of the deep waters even at Station "P" in the North Pacific Ocean (Parsons and Strickland 1962). It is likely that particulate matter in the continental shelf area is more affected by these terrigenous cellulose materials than that in deep water. However, with the present state of knowledge, the source of the glucose cannot be clearly determined because of the limited information available on the chemistry of polysaccharides of particulate matter.

Fatty acids

Intensive studies have been conducted to analyze for fatty acids in all kinds of biological materials found in the ocean in order to determine the nutritional value of various items in the marine food chain. In this study, emphasis was placed on determining the fatty acid content of particulate matter found in the Bering Sea and evaluating the biological activity of these particulate materials, especially those collected from the surface and subsurface waters. Some effort is also given to the utilization of fatty acid distribution data in vertical transport studies.

Particulate samples were collected from station 4 of the cruise of KJ-78-3 in the Bering Sea. After conversion of fatty acids extracted from the particulate matter to corresponding methyl esters, the methyl ester composition was determined by gas chromatography or by combined gas chromatography and mass spectrometry. The fatty acid composition of the particulate matter at station 4 of KH-78, shown in Table 23-5, ranged from 1.4 to 6.3 $\mu\text{gC/l}$, which accounted for 4.9-11.7 percent of the total organic carbon. The concentration tended to decrease with depth while the ratio of fatty acid carbon to particulate organic carbon tended to increase with depth.

High proportions of $\text{C}_{14:0}$, $\text{C}_{16:0}$, and $\text{C}_{16:1}$ acids to total fatty acids were found in the particulate samples from surface and subsurface waters. The particulate matter collected here consisted mainly of diatoms (75 percent in cell number) with some dinoflagellates (25 percent). This distribution agrees well with the fact that the Bacillariophyceae have been distinguished from other classes of phytoplankton having a high level of $\text{C}_{14:0}$ and $\text{C}_{16:0}$ acids, but a low level of $\text{C}_{18:0}$ acid (Ackman et al. 1968).

One of the characteristic features of the particulate matter from station 4 is the high values of poly-

TABLE 23-5

Fatty acid compositions of particulate matter at station 4 of KH-78-3.

Fatty acid		Depth (m)					
		1	50	110	177	500	2,000
iso ^a	14:0 ^b	tr ^c	0.6	2.2	1.2	0.5	0.6
	14:1	tr	1.0	1.3	0.9	0.4	0.4
	14:0	17.7	13.7	10.1	9.6	6.7	7.4
anteiso ^a	15:0	0.3	1.9	1.5	1.5	0.8	0.7
iso	15:0	1.3	3.1	3.0	2.4	1.2	1.3
	15:0	0.9	2.7	6.0	4.5	2.9	3.6
	16:2, 16:3, 16:4	3.0	tr	tr	1.1	ND	ND
	16:1	12.3	15.5	12.4	8.0	4.4	4.8
	16:0	26.7	41.4	31.1	36.2	34.0	38.4
anteiso	17:0	0.8	0.9	2.2	2.2	1.0	1.1
iso	17:0	0.5	1.1	2.2	1.4	0.9	0.9
	17:0	0.4	0.7	1.5	1.4	1.8	1.8
	18:2, 18:3, 18:4	13.5	3.8	tr	1.6	tr	1.6
	18:1	17.8	10.5	8.8	6.3	4.4	4.3
	18:0	5.0	1.3	13.0	18.7	37.9	27.5
	19:0	tr	0.8	0.8	0.3	0.4	0.4
	20:0	tr	0.6	0.9	0.9	1.2	1.5
	22:0	0.8	0.4	1.6	0.5	0.8	0.8
	23:0	tr	tr	tr	0.2	tr	0.3
	24:0	tr	tr	1.3	0.8	0.8	1.5
	25:0	tr	tr	tr	tr	tr	0.3
	26:0	tr	ND ^d	tr	0.2	tr	0.9
	27:0	ND	ND	ND	ND	ND	ND
	28:0	ND	ND	tr	tr	tr	tr
Total ($\mu\text{gC/l}$)		6.3	3.4	1.4	2.9	2.6	3.3
FA-C/OC ^e (%)		4.9	5.8	6.6	12.0	11.7	10.5
N/B ^f		34.5	14.2	11.2	13.3	25.6	25.0

^aBranched fatty acids^bAll fatty acids are given by normal chain length: number of double bond.^cTrace^dNone detected^eFatty acid carbon/organic carbon^fNormal acid/iso- and anteiso-acid

unsaturated C_{18} acids, which are important biochemically in connection with the photosynthetic processes of diatoms—for example, in the stimulation of chloroplast formation (Rosenberg and Gouanx 1967) and as an active intermediate of photosynthetic carbon transfer (Kates and Volcani 1966). Schultz and Quinn (1977) observed that in coastal waters the total concentration of fatty acids increased as did the relative concentration of polyunsaturated C_{18} acids as the phytoplankton bloom progressed. The polyunsaturated acids varied from 7 percent at the start to 12 percent at the peak of the phytoplank-

ton bloom, then dropped to 3 percent at the end of the bloom in coastal waters. These data are compatible with the work of Jeffries (1970), who found that the $\text{C}_{18:2}$ acid accounted for 4 percent and 7 percent of total fatty acids in the summer-fall and winter-spring phytoplankton assemblages respectively.

From the data in Table 23-5, it may be deduced that phytoplankton of the particulate matter from station 4 were actively growing cells sampled during the period of the phytoplankton bloom. Of the total fatty acids of the surface particulate matter, approximately 30 percent were found to be unsaturated C_{18}

acids, which concentration decreases rapidly with depth. These data, coupled with the literature cited above, point to the possible utility of this kind of observation as a diagnostic tool in assessing the biological activity of phytoplankton in marine environments.

Several branched (iso- and anteiso-) fatty acids of C_{14} , C_{15} , and C_{17} were found in the particulate samples from the surface through deep waters with lower values than those obtained in the sedimentary samples from various sources. Extensive work by Leo and Parker (1966) and Cooper and Blumer (1968) indicates that iso- and anteiso-acids are potential indicators of bacterial contribution to sedimentary organic matter because branched acids predominate over the straight chain acids in bacteria (Kates 1964, Kaneda 1967). Blumer (1970) reported that the ratio of normal acid to iso- and anteiso-acids was calculated to be 3-7 and 50-100 for marine sediments and marine plankton, respectively, and low values of the ratio in the sediments were due to bacterial contamination. In view of these data, the particulate matter from the surface water (1 m) sampled at station 4 was almost free from bacterial contamination as indicated by the relatively high value (34.5) of the normal acid/iso- and anteiso-acids observed. Bacterial contamination of the particulate matter is, however, evident in the intermediate waters from 50 to 177 m depth, where low values of the normal acid to branched acid ratio were found. Such phenomena may be expected to occur in the intermediate waters since the particulate organic matter is available to bacteria once it sinks from the euphotic zone to the underlying intermediate waters. The selective degradation of normal acids, especially unsaturated acids, may also affect the acid/iso- and anteiso-acid ratios to some extent.

All of the fatty acid components decrease in abundance in the particulate matter with depth except the saturated C_{18} acid, which increased from $0.32 \mu\text{gC/l}$ at 1 m to $0.91 \mu\text{gC/l}$ at 2,000 m. Since saturated C_{18} acid is only a minor component of the fatty acids of phytoplankton, zooplankton (Hinchcliffe and Riley 1972, Lee et al. 1971), and fishes (Bishop et al. 1976), these biological materials are not likely to be a direct source of this accumulation. Selective degradation of the particulate fatty acids is a more likely source of this material, but more complex processes may cause the increment increase in the particulate matter found at depth. Sorption of dissolved fatty acids by mineral particles in sea water may be another process functioning to increase the concentration of saturated C_{18} acid in the deep waters. The sorption process is controlled by physi-

co-chemical factors (temperature, pH, and salinity) (Meyers and Quinn 1973), the carbon number of fatty acids (Barcelona and Atwood 1979), and the size of the mineral particles (Morris and Calvert 1975). The observation that saturated C_{18} acid is the most abundant of dissolved fatty acids (Williams 1965, Ackman and Hooper 1970) in suspended particles of 4-10 nm in diameter (Tsunogai et al. 1977) suggests that the sorption processes of dissolved fatty acids with mineral particles may be functional in causing the increased concentration observed.

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Hydrocarbons of Animals of the Bering Sea

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ABSTRACT

Samples of biological materials including pelagic animals and marine birds and mammals have been collected in the Bering Sea for hydrocarbon determination. Pristane, a compound derived from chlorophyll of zooplankton, was both abundant and ubiquitous, being detected in 91 percent of the tissues analyzed in concentrations up to 220 $\mu\text{g/g}$. Heneicosahexaene was also common, detected in 32 percent of the samples in concentrations up to 88 $\mu\text{g/g}$. However, the latter compound, which is produced by marine algae, was essentially confined to organisms low in the food web. No hydrocarbons from petroleum or terrigenous plant sources were detected in any of the animal tissues analyzed. Our analyses show that the source of hydrocarbons in the Bering Sea pelagic environment is biosynthesis in that environment. This is in keeping with the current understanding of productivity and carbon flow in this area.

INTRODUCTION

In many coastal environments the hydrocarbon constituents of marine animals include important contributions from anthropogenic and natural terrigenous sources. However, all of the hydrocarbons identified in organisms collected in the southeastern Bering Sea in the spring of 1976 and 1977 appear to have had their origin in the marine pelagic system of that region. We base this conclusion on the results of the analyses of 34 samples of plankton, fish, bird, and marine mammal tissues carried out as part of a survey of the kinds and amounts of ambient hydrocarbons in the Bering Sea environment. These are the first reported determinations of hydrocarbons in a suite of biological materials collected in the Bering Sea.

METHODS

Plankton, fish, and bird samples which were collected in 1977 expressly for these hydrocarbon

analyses were immediately frozen in specially cleaned (500 C, 24 hours) glass containers. Marine mammal tissue samples had been collected in 1976 as part of a study of general biology and life history of Alaskan marine mammals. These tissues had been wrapped in aluminum foil, placed in plastic bags, and frozen.

In the laboratory approximately 10 g of tissue was extracted for two hours at 90 C in a capped centrifuge tube containing 10 ml 4 N aqueous KOH and 2 ml hexane. After the extract had cooled to room temperature, an additional 10 ml of hexane was added. The tube was then shaken well and centrifuged for 15 minutes at 2,500 rpm. The hexane (upper) phase was removed by pipette and the aqueous phase extracted twice more with hexane in the same manner. The hexane extracts were combined and dried over anhydrous sodium sulfate. The hexane was concentrated to about 5 ml and an aliquot evaporated and weighed to determine the total nonsaponifiable lipids. Extracts were column chromatographed on silica gel (5 percent water). Saturated hydrocarbons were eluted first with hexane. Next unsaturated hydrocarbons were eluted with 40 percent benzene in hexane. Eluates were concentrated to approximately 1 ml for analysis by gas chromatography (GC). Each fraction was first analyzed by flame ionization GC (Hewlett Packard 5710) using a 50 m by 0.7 mm support coated open tubular column with OV-101 serving as stationary phase. Quantification was accomplished by a digital integrator (Hewlett Packard 3380 or 3385) and corrected for percentage recovery. Hydrocarbons were identified through the use of internal and external standards and by mass spectrometry (MS)

using a Hewlett Packard 5390/5933 computerized GCMS system.

RESULTS

Table 24-1 shows the concentrations of hydrocarbons found in Bering Sea animals; Table 24-2 gives the dates and locations of sample collections. Seventeen samples of planktonic invertebrate material were analyzed from the following taxa: *Neomysis rayi* (a mysid), *Parathemisto pacifica* (an amphipod), Medusae (jellyfish), Chaetognatha (arrow worms), and Euphausiacea (euphausiids). The mysid sample was essentially devoid of hydrocarbons. However, the remainder of this group all contained substantial concentrations of pristane and several also contained heneicosahexaene (21:6). One sample of fish, *Malotus villosus* (the common capelin), was analyzed. Its principal hydrocarbon constituent was pristane. Livers from three species of sea birds were also investigated. These included *Uria aalge* (Common Murre), *Uria lomvia* (Thick-billed Murre), and *Rissa tridactyla* (Black-legged Kittiwake). The murre livers were quite low in hydrocarbons, but the kittiwake livers had a very high concentration of pristane. Samples of kidney were analyzed from four species of seal, *Erignathus barbatus* (bearded seal), *Histiophoca fasciata* (ribbon seal), *Phoca vitulina richardsi* (harbor seal), and *P. vitulina largha* (spotted seal). These species showed low to moderate concentrations of hydrocarbons, including pristane. Additional spotted seal tissues, including liver, muscle, and blubber, were also studied. While the hydrocarbon contents of these tissues were qualitatively similar to those of the kidneys examined, the liver and blubber tended to have substantially higher hydrocarbon concentrations. Ten of the thirteen seal tissues investigated also contained phthalate esters, synthetic organic compounds used in the manufacture of some plastics.

DISCUSSION

Pristane, which was found in all but three of the samples analyzed in this study, is a well-known marine biogenic hydrocarbon. Pristane has been identified in numerous species of zooplankton and is particularly abundant in *Calanus* (Blumer et al. 1963). Avigan and Blumer (1968) demonstrated that pristane is derived in calanoids from dietary chlorophyll. Blumer (1967) has also shown that pristane can be retained by higher marine organisms which feed on zooplankton. A sizable body of subsequent work has confirmed the large extent to which pristane is produced and transferred without chemical

alteration by marine animals (National Academy of Sciences 1975). Our finding that pristane is both ubiquitous and abundant in Bering Sea biota is further confirmation of this.

The three highest concentrations of pristane, 220, 104, and 101 $\mu\text{g/g}$, were observed in the liver of the black-legged kittiwake and in two samples of spotted seal blubber. This finding is consistent with the idea (Teal 1977) that higher marine animals without gills (birds and mammals) may accumulate higher hydrocarbon concentrations since they lack the efficient mechanism that gills afford for partitioning lipids into seawater. However, it must be noted that the highest concentration of pristane observed, 220 $\mu\text{g/g}$ in the kittiwake, is only a factor of three higher than the highest observed in an invertebrate.

In the tissues of spotted seal which were examined individually, pristane concentration tends to be high in the blubber, intermediate in the liver, and low in muscle and kidney, consistent with general solubility. But for individual seals there are exceptions to this trend. For instance, in animal 30-76 the lowest pristane concentration is found in the liver, while in animal 20-76 the concentration in the liver is higher than in the blubber. These apparent anomalies may reflect the different hydrocarbon turnover times of the tissues and the nutritional states of the individual seals.

Heneicosahexaene, the all *cis*-3,6,9,12,15,18 isomer (21:6), has been identified in marine phytoplankton by Blumer et al. (1970) and independently by Lee et al. (1970). They showed that considerable variation exists in the concentrations of this compound produced by various species of algae and also in the extent to which 21:6 is accumulated by various species of zooplankton fed on phytoplankton rich in this compound. Subsequent work has shown that 21:6 occurs widely in marine algae (Blumer et al. 1971, Youngblood and Blumer 1973). Because 21:6 is labile chemically (Blumer et al. 1970) and undoubtedly also biochemically, it is not accumulated by consumer organisms nearly to the extent that pristane is. Our finding that among the animals of the Bering Sea 21:6 is largely restricted to the zooplankton is consistent with these generalizations.

Most of the seal tissues analyzed contained phthalate esters. Phthalates have been reported in marine water, sediment, air, and biota (Giam et al. 1978); however, we cannot exclude the possibility that the phthalates in these tissues were introduced as contaminants after sample collection. These seal samples were not collected for hydrocarbon determination; rather they were collected and saved as part of a study of the general biology and life history of

TABLE 24-1

Hydrocarbon concentrations ($\mu\text{g/g}$, Wet Weight) in Bering Sea animals

Material	Sample number	Total saturated	Total unsaturated	Pristane	21:6	Phthalate
<i>Neomysis rayi</i>	38-50	< 0.01	< 0.01	—	—	—
<i>Parathemisto pacifica</i>	50-60	70	20	70	—	—
<i>Parathemisto pacifica</i>	21-36	68	4.1	68	3.4	—
<i>Parathemisto pacifica</i>	21-37	20	13	20	—	—
Medusae	50-62	2.3	7.5	2.3	1.2	—
Chaetognatha	11-28	8.3	0.9	8.3	0.9	—
Chaetognatha	2-4	60	10.3	60	9.0	—
Chaetognatha	5-18	53	2.3	53	2.3	—
Chaetognatha	7-24	34	8.9	34	7.6	—
Chaetognatha	11-29	7.5	0.6	7.5	—	—
Chaetognatha	11-30	11.1	< 0.01	11.1	—	—
Euphausiacea	2-1	52	31	52	27	—
Euphausiacea	7-21	66	150	58	—	—
Euphausiacea	7-23	58	100	49	88	—
Euphausiacea	21-38	9.5	34	5.7	26	—
Euphausiacea	44-52	3.6	34	1.2	—	—
Euphausiacea	50-61	1.0	11	1.0	10.9	—
<i>Mallotus villosus</i>	38	10	1.1	10	—	—
<i>Uria aalge</i>	77-30	< 0.01	< 0.01	—	—	—
<i>Uria lomvia</i>	77-37	0.4	0.6	0.4	—	—
<i>Rissa tridactyla</i>	77-31	220	0.8	220	—	—
<i>Erignathus barbatus</i> kidney	22-76	< 0.01	0.25	—	—	—
<i>Histiophoca fasciata</i> kidney	24-76	0.21	—	0.21	—	0.88
<i>Phoca vitulina richardsi</i> kidney	12-76	0.95	—	0.52	—	1.1
<i>Phoca vitulina largha</i> kidney	30-76	4.5	—	4.5	—	5.5
muscle	30-76	0.85	—	0.85	—	20
liver	15-76-1	66	—	66	—	18
liver	15-76-2	65	1.7	65	—	9.3
liver	20-76-1	30	2	30	0.1	14
liver	20-76-2	43	—	43	—	19
liver	30-76	0.22	—	0.22	—	2.2
blubber	15-76	104	—	104	—	—
blubber	20-76	23	140	23	—	—
blubber	30-76	140	—	101	—	37

TABLE 24-2

Date and location of collection of samples for hydrocarbon determination

Material	Sample Number	Date	Latitude N	Longitude W
<i>Neomysis rayi</i>	38-50	2 June 1977	60°27'	170°01'
<i>Parathemisto pacifica</i>	50-60	8 June 1977	57°59'	168°49'
<i>Parathemisto pacifica</i>	21-36	28 May 1977	59°31'	174°48'
<i>Parathemisto pacifica</i>	21-37	28 May 1977	59°31'	174°48'
Medusae	50-62	8 June 1977	57°59'	168°49'
Chaetognatha	11-28	23 May 1977	58°39'	172°15'
Chaetognatha	2-4	22 May 1977	54°42'	165°59'
Chaetognatha	5-18	22 May 1977	55°33'	168°09'
Chaetognatha	7-24	23 May 1977	56°05'	169°40'
Chaetognatha	11-29	23 May 1977	58°39'	172°15'
Chaetognatha	11-30	23 May 1977	58°39'	172°15'
Euphausiacea	2-1	22 May 1977	54°42'	165°59'
Euphausiacea	7-21	23 May 1977	56°05'	169°40'
Euphausiacea	7-23	23 May 1977	56°05'	169°40'
Euphausiacea	21-38	28 May 1977	59°31'	174°48'
Euphausiacea	44-52	6 June 1977	60°22'	169°07'
Euphausiacea	50-61	8 June 1977	57°59'	168°49'
<i>Mallotus villosus</i>	38	2 June 1977	60°27'	170°01'
<i>Uria aalge</i>	77-30	25 May 1977	60°37'	174°38'
<i>Uria lomvia</i>	77-37	25 May 1977	60°37'	174°38'
<i>Rissa tridactyla</i>	77-31	25 May 1977	60°37'	174°38'
<i>Erignathus barbatus</i>	22-76	28 March 1976	56°12'	165°29'
<i>Histiophoca fasciata</i>	24-76	19 April 1976	57°21'	172°41'
<i>Phoca vitulina richardsi</i>	12-76	25 March 1976	56°08'	164°20'
<i>Phoca vitulina largha</i>	15-76	26 March 1976	56°05'	164°31'
<i>Phoca vitulina largha</i>	20-76	27 March 1976	56°05'	164°31'
<i>Phoca vitulina largha</i>	30-76	24 April 1976	56°05'	162°47'

Alaskan marine mammals. The samples had been wrapped in aluminum foil, placed in plastic bags (a potential contamination source), and frozen. But we tend to believe that post-collection contamination is not the only source of these phthalates, since the smaller samples analyzed did not show the higher concentrations one might expect if various-sized pieces of tissue were contaminated from a constant source such as an outer plastic wrapper. Moreover,

seals have been shown to accumulate chlorinated hydrocarbons (Holden and Marsden 1967).

Considerable significance also lies in the groups of hydrocarbons which were not found in this suite of samples. The hydrocarbons associated with higher terrigenous plants were not observed. This group of hydrocarbons would include odd chain length normal alkanes with 23 to 31 carbon atoms. Their absence indicates that, at least in the spring, terrigenous

hydrocarbon sources are minor compared to marine sources, in accord with the well-known high primary productivity of the Bering Sea. But this relationship may not continue throughout the year, since marine primary production is highly seasonal in northern waters. Shaw and Baker (1978) have documented that at roughly the same latitude, in Port Valdez, Alaska, pristane dominates the hydrocarbons of intertidal biota during the summer but terrigenous normal alkanes dominate in winter.

Another group of hydrocarbons not observed in the animals of the southeastern Bering Sea is the fossil hydrocarbons. This indicates that neither natural petroleum seepage nor pollution is resulting in significant accumulations of petroleum hydrocarbons in marine animals of the area. It should be noted, however, that the analytical techniques used in this study would not have detected low levels of poly-

cyclic aromatic hydrocarbons. Since evidence has been presented that atmospheric transport has brought this class of compounds to the sediments of the Gulf of Alaska (Laflamme and Hites 1978) and the arctic (Shaw et al. 1979), their presence in the Bering Sea in trace amounts cannot be excluded.

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Organic Geochemistry of Surficial Sediments from the Eastern Bering Sea

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ABSTRACT

The distribution and concentration of hydrocarbons in surficial sediments from the continental shelf of the eastern Bering Sea were determined as part of an environmental survey. Gravimetric and gas chromatographic analyses of the aliphatic fractions indicate that the hydrocarbons are predominantly allochthonous detritus, probably transported by rivers, with minor autochthonous elements. Gas chromatographic/mass spectrometric characterization of aromatic fractions indicates that the source of aromatic compounds may be pyrolytic.

The fact that autochthonous hydrocarbons in the sediments from this region noted for its high biological productivity are found in relatively small amounts suggests that rapid and efficient recycling of marine lipids has occurred within the water column or at the sediment-water interface.

Comparison of hydrocarbon distribution patterns of open-shelf sediments with those of eelgrass and sediments from Izembek Lagoon indicates that coastal lagoons are probably not a major source of hydrocarbons in the southeastern Bering Sea sediments. However, carbon isotopic composition of humic acids and protokerogens isolated from the lagoon and shelf sediments suggests a possible source relationship.

It is possible that the terrigenous hydrocarbons in the Norton Sound region come mainly from the Yukon River. Nine samples from Norton Sound collected over a period of three years near suspected petroleum seepage show alkane distribution patterns which are not characteristic of weathered petroleum.

Comparison of data from the Bering Sea region with those from the Southern California Bight, known to be petroleum-contaminated, clearly shows that the Alaskan sediments are "clean" and generally free of petrogenic material.

INTRODUCTION

The eastern Bering Sea is noted for its high biological productivity and renewable biological resources. It is also a potentially important petroleum-producing area. And yet little is known about the distribution of organic matter in its sediments. In general, organic

geochemical data from the Alaskan outer continental shelf are relatively scant except for the hydrocarbon studies of Peake et al. (1972), Kaplan et al. (1977, 1979) in the Gulf of Alaska and Bering and Beaufort seas, Shaw et al. (1978) on Beaufort Sea sediments, Chester et al. (1976) on sediments from Prince William Sound, the northeast Gulf of Alaska, Kinney (1973) in Cook Inlet and Port Valdez, and Wong et al. (1976) in the Beaufort Sea, along the Canadian border. In our laboratory, we have undertaken a study of hydrocarbon distribution and concentrations in the surficial sediments of the eastern Bering Sea. The analyses provide details of the natural hydrocarbon background for the study of marine processes and for monitoring hydrocarbon pollution in conjunction with an increase in offshore petroleum production. In this chapter, we present results of the analyses of the aliphatic and aromatic fractions (along with other supplementary geochemical data) of approximately 20 sediment samples from the southeastern Bering Sea area, a sample of eelgrass and the sediment within the eelgrass environment of Izembek Lagoon on the northwest side of the Alaska Peninsula, and about 40 sediment samples from the Norton Sound region (northeastern Bering Sea).

EXPERIMENTAL PROCEDURES

Sampling

The locations of the sampling stations are given in Figs. 25-1a and b. The surface sediments were collected with steel Van Veen grab sampler during the summer 1975 cruises of the NOAA ship *Discoverer*,

Figure 25-1a. Location of southeastern Bering Sea stations.

Leg III, and with modified aluminum Van Veen grab sampler (Soutar 1976) in the 1976 and 1977 cruises of the R/V *Sea Sounder*. About 20 sediment samples, including two box-core samples, were collected in the summer 1979 cruise of the *Discoverer*. Care was taken during sampling to avoid contamination from the sampler or the ship. Except for a hinge in the sampler that could not be easily cleaned, all materials that came in contact with the sediment were cleaned and rinsed with organic solvents. Sediments were placed in prewashed glass jars and frozen until analysis.

Carbon and sulfur analysis

Elemental analysis of sulfur was carried out on freeze-dried sediment samples combusted in a LECO (Laboratory Equipment Corporation) Model No. 523 induction furnace and the resulting sulfur gases were titrated according to ASTM procedure E30-47, using a LECO Model No. 517 titrator. Total carbon and organic carbon (carbon remaining after treatment with 3N HCl) were determined by combustion of dry

sediment samples in a LECO Model No. 572-100 semiautomatic, acid-base carbon determinator. Details of these procedures are given in the pertinent instruction manuals of LECO.

Extraction of hydrocarbons

The frozen sediment (southeastern Bering Sea) was placed in precleaned cellulose extraction thimbles and washed with distilled water to remove salts. The wet sediment was freeze-dried for 48 hours and extracted in a Soxhlet extractor for 100 hours with toluene:methanol (3:7) with one solvent change after 24 hours. Hexane extract of the water wash was combined with these extracts, reduced to a small volume on a rotary evaporator at 38 C, and treated with activated copper to remove sulfur (MacLeod et al. 1976). The extract was saponified by refluxing for four hours with 0.5 N KOH in a 1:1 mixture of water and methanol; the nonsaponifiable portion was then extracted into hexane and fractionated on a glass column packed with silica gel beneath neutral alumina (activity grade 1). A column with a length-to-

Figure 25-1b. Locations of Norton Sound stations. ●1976 stations; + 1977 stations; ▲1979 stations.

internal diameter (I.D.) ratio of 20:1 was packed with a weight ratio of 100 parts alumina and 200 parts silica to one part sample. Elution with two column volumes each of hexane, benzene, and methanol was carried out. The benzene fractions were further purified by thin-layer chromatography on silica gel plates developed in CH_2Cl_2 to remove the methyl esters. The gas chromatographic analysis of the hexane fractions was carried out using a Hewlett-Packard Model No. 5830A instrument, equipped

with a linear temperature programmer, FID detector, and electronic integrator. A glass SCOT column, 50 m $\times$ 0.4 mm I.D., coated with OV-101 (SGE Scientific, Inc.) was used. The column was temperature-programmed from 100 C to 275 C at 2 C/min. with helium carrier gas flow of 4 ml/min. and held isothermally for 60 min.

Norton Sound sediment samples were extracted with methanol for 24 hours to avoid freeze-drying, then with toluene:methanol for 76 hours and proc-

essed as discussed earlier for column chromatography after partitioning the methanol extract with hexane. Only silica gel was used for column fractionation and aliphatic and aromatic fractions free of methyl esters were eluted (Venkatesan et al. 1980). Glass capillary column coated with OV-101 (J&W), 30 m long and 0.25 mm I.D. was used in the Hewlett-Packard Model 5840A gas chromatograph. It was programmed from 35 C to 260 C at 4 C/min and then held isothermal for two hours. The flow rate of helium carrier gas was 3.6 ml/min.

Selected samples were analyzed by gas chromatography/mass spectrometry on a Finnigan model 4000 quadrupole mass spectrometer directly interfaced with a Finnigan Model 9610 gas chromatograph. The GC was equipped with a glass capillary column like the one described above for the analysis of aliphatic fractions. The aromatic fractions were analyzed using an SE 54 (J&W) column. The mass spectrometric data were acquired and processed using a Finnigan Incos Model 2300 data system.

Solvents were of high purity grade (Burdick and Jackson "distilled in glass" grade). Trace organic compounds in double-distilled water were removed by passing through a column of Chromosorb 102. NaCl was heated at 500 C overnight and KOH fused at 500 C for two hours. Silica gel and alumina were sonicated twice with methylene chloride-methanol mixture and once with hexane prior to activation.

RESULTS

The eastern Bering Sea shelf has been arbitrarily divided in this study into two regions, southeastern Bering Sea and Norton Sound (northeastern area), for the sake of convenience in discussing the data. Results of analysis of 8 samples from about 20 samples collected in summer of 1979 from Norton Sound are also included. The range of organic carbon (0.14-1.3 percent) given in Tables 25-1 and 25-2, and sulfur (0.014-0.1 percent) are low compared to many fine-grained Recent marine sediments (Degens 1967, Didyk et al. 1978). Sample 48 from Norton Sound collected in 1977 has an anomalous organic carbon content (4.23 percent) and the total carbon content is also unusually high (9 percent). However, a sample collected very close to this station (160) in 1976 had a more normal carbon content of 0.7 percent. In summary, the data are typical of other unpolluted, relatively coarse marine sediments. The total hydrocarbons (Tables 25-1 and 25-2) range from 1 to 29 $\mu\text{g/g}$ of dry sediment with the exception of EBBS 35, which has a value of 241 $\mu\text{g/g}$. This range of values is low compared to unpolluted Recent

marine sediments from other environments (30 and 100 ppm) and is 10-100 times lower than the values of total hydrocarbon in sediments contaminated by petroleum (Palacas et al. 1976, Farrington and Tripp 1977, Crisp et al. 1979, Venkatesan et al. 1980).

Representative histograms of alkane distributions and gas chromatographic traces of a few of the hexane fractions are illustrated in Figs. 25-2, 25-3, and 25-4. The range in total *n*-alkanes resolved by gas chromatography is from 0.1 to 9 $\mu\text{g/g}$ dry sediment and reflects the variability in sediment grain size, distance from the source, and organic carbon content. Although the chromatograms differ in detail, the hydrocarbon distributions are similar in many respects. An *n*-alkane series with carbon numbers of 16-34 is present in all sampling areas corresponding to those generally reported for marine sediments. The pronounced odd carbon preference (Tables 25-1 and 25-2) with *n*-C₂₇ or *n*-C₂₉ predominating (Fig. 25-3), is not characteristic of *n*-alkanes in petroleum. Pristane, phytane, and unidentified branched or unsaturated components are present generally in low amounts.

Gas chromatographic traces of hexane fraction in eelgrass and sediment from Izembek Lagoon and isotopic analysis data on humic substance and proto-kerogens are presented in Fig. 25-5 and Table 25-3, respectively.

Multiple homologous series of extended diterpanes and triterpanes were identified by GC/MS analyses and relative distribution histograms of a few stations are shown in Fig. 25-6. An example of a histogram of a sample from Southern California Bight is included for comparison.

Concentrations of specific aromatic compounds in samples analyzed by GC/MS are given in Table 25-4.

DISCUSSION

Elemental analysis

The organic carbon values are low considering the relatively high biological productivity of this continental shelf region (Simoneit 1975), apparently as a result of a combination of oxic conditions at the sediment surface and a high-energy depositional environment (Sharma 1974).

When the sediment type in samples from the southeastern Bering Sea is classified according to grain size (sand, silt, and clay) after Shephard (1954), it is clear that sediments having lower organic carbon (0.3-0.4 percent) are sandy and those with higher organic carbon content (0.6-0.8 percent) are either silty or rich in silt-clay. Thus, the distribution of organic carbon here supports the findings of Bordovskiy

Table 25-1.
Gravimetric and gas chromatographic data of southeastern
Bering Sea sediment samples

Station number*	Aliphatic fraction ($\mu\text{g/g}$)	Aromatic fraction ($\mu\text{g/g}$)	n-alkanes ($\mu\text{g/g}$)	Organic carbon (%)	HC $\times 10^4$ OC	n-alkanes $\times 10^4$ OC	Pr Ph	Odd Even
8	5.7	2.8	0.56	0.23	36.9	2.4	2.70	2.99
12	3.4	1.4	0.33	0.14	34.3	2.3	3.32	1.76
17	13.0	5.2	1.09	0.76	23.9	1.4	3.97	3.43
19	7.4	4.5	2.57	0.39	30.5	6.6	5.81	3.17
24	6.1	5.4	0.66	0.33	34.8	2.0	3.39	2.96
28	8.7	4.1	2.93	0.59	21.7	5.0	10.20	4.09
35	180.1	60.8	n.r.	0.41	587.6	n.r.	n.r.	n.r.
37	5.8	4.0	0.76	0.41	23.9	1.8	1.76	3.28
38	4.9	10.6	1.64	0.66	23.5	2.5	5.18	4.41
40	1.9	2.6	0.61	0.32	14.1	1.9	3.37	3.41
41	1.4	0.5	0.41	0.37	5.1	1.1	n.d.	3.80
43	2.4	2.7	0.52	0.30	17.0	1.8	2.26	3.08
46	4.3	7.5	0.74	0.42	28.1	1.8	17.90	3.59
51	2.8	0.6	0.77	n.d.	25.4	n.d.	3.49	2.56
54	7.4	9.9	2.10	0.68	25.4	3.1	8.80	2.57
56	10.6	8.5	0.75	0.47	40.6	1.6	2.93	3.78
58	3.8	2.8	0.28	0.31	21.3	0.9	4.80	3.33
59	6.4	6.2	1.55	0.27	46.7	5.7	1.74	2.85
64	12.3	9.8	1.79	0.77	28.7	2.3	2.27	2.75
65	6.9	9.6	1.60	0.67	24.6	2.3	16.43	3.77
45B	3.9	4.9	0.78	0.76	11.6	1.0	3.78	1.96

*Bulk samples of the upper 0-10 cm of surface sediment

Aliphatic = eluted by hexane

Aromatic = eluted by benzene and then cleaned by TLC procedure to remove methyl esters

HC = total hydrocarbons, sum of aliphatic and aromatic fractions in $\mu\text{g/g}$ dry sediment

n-alkanes = resolved by gas chromatography

Pr = Pristane

Ph = Phytane

Odd/Even = Summed from C_{15} to C_{34}

n.r. = not resolved

n.d. = not determined

TABLE 25-2

Gravimetric and gas chromatographic data of Norton Sound sediment samples

Station No.	Aliphatic fraction ($\mu\text{g/g}$)	Aromatic fraction ($\mu\text{g/g}$)	<i>n</i> -alkanes ($\mu\text{g/g}$)	Organic carbon (%)	$\frac{\text{HC}}{\text{OC}} \times 10^4$	$\frac{n\text{-alkanes}}{\text{OC}} \times 10^4$	$\frac{\text{Pr}}{\text{Ph}}$	$\frac{\text{Odd}}{\text{Even}}$
1976								
47	9.6	7.5	3.28	0.93	18.4	3.7	2.00*	5.38
49	24.8	4.1	5.69	1.12	25.8	5.1	1.50*	6.06
70	2.2	6.2	0.01	0.31	27.1	0.1	8.00	1.65
88B	3.9	5.7	0.69	0.53	18.2	1.3	2.14	4.11
105	1.8	0.9	0.07	0.93	2.9	0.1	n.d.	11.21
125	0.1	2.4	0.69	1.18	2.6	1.3	2.00	4.02
131	9.0	2.9	7.18	0.96	27.3	6.3	n.d.	2.80
137	17.8	4.5	8.69	n.d.	n.d.	n.d.	3.00*	4.07
147	6.8	2.3	2.24	0.33	27.5	6.8	1.80*	2.85
154	16.3	4.2	5.45	0.99	20.7	5.5	3.60*	5.69
156	7.1	5.5	5.06	1.30	9.7	3.9	2.67*	5.57
162	2.3	2.3	0.45	0.92	5.0	0.5	2.00*	4.75
166S	1.1	0.8	0.16	1.16	1.6	0.1	4.00*	5.16
168S	3.2	2.2	1.48	1.10	4.9	1.4	3.50*	5.26
169S	2.6	4.0	0.95	0.33	20.1	2.9	3.60	4.47
170S	4.4	2.2	2.57	0.52	12.8	4.9	n.d.	5.80
172S	10.9	3.8	2.89	0.87	16.9	3.3	n.d.	4.70
174S	3.9	2.0	1.79	0.82	7.2	2.2	6.00*	4.50
1977								
34	0.8	0.7	0.09	0.12	12.4	0.8	2.0*	4.55
35	2.2	1.1	0.57	0.59	5.7	0.9	7.0*	5.15
39S	0.6	0.2	0.09	0.38	2.3	0.2	2.5*	5.35
41S	2.5	0.8	0.23	0.44	7.5	0.5	4.0*	4.78
42S	4.4	1.9	0.83	0.32	19.9	2.6	5.5*	4.78
43	1.0	1.7	0.38	0.60	4.5	0.6	6.5*	4.22
44	2.1	0.9	0.25	0.52	5.7	0.5	6.0*	3.21
48S	5.8	5.0	1.60	4.23	2.6	0.4	5.0*	6.37
14 π	5.4	1.3	1.22	0.28	23.8	4.4	3.1*	5.26
17 π	5.5	2.7	1.75	0.24	18.9	2.0	1.3*	5.12
17†	14.1	2.2	2.18	0.86	26.3	6.4	3.0*	5.67
17§	3.2	0.9	0.95	0.50	8.0	1.9	4.0*	5.34
1979								
13	4.5	5.1	2.08	0.38	25.2	5.4	1.65	4.32
15	4.0	1.8	1.63	0.48	11.9	3.4	2.47	3.81
18	3.2	1.5	2.22	0.48	9.8	4.6	1.86	4.04
20	5.4	1.5	1.50	0.40	17.3	3.8	n.d.	4.03
22	8.9	5.5	2.83	0.86	16.8	3.3	2.44	3.35
25	5.2	1.9	1.57	0.66	10.8	2.4	n.d.	4.11
33A	1.7	0.9	0.29	n.d.	n.d.	n.d.	n.d.	4.09
47A	1.4	1.1	0.44	n.d.	n.d.	n.d.	n.d.	4.29

Aliphatic = eluted by hexane; Aromatic = eluted by hexane:benzene (3:2 by volume); * = approximate values based on peak heights. The rest are the same as those given in Table 25-1.

Samples are 0-2 cm except B = bulk; S = surface; π = 0-3 cm; † = vibracore, 0-3 cm; § = 160 cm vibracore. Samples 14-17 belong to a different program (USGS).

(1965) and Sharma (1974), that organic carbon increases with decreasing mean grain size of the sediment. Sediments in this region are reported to become progressively finer grained from nearshore to the edge of the shelf (Sharma 1974). The total hydrocarbon content of these sediments follows the same trend, with low concentrations of total hydrocarbons in coarse-grained sediments close to shore and higher concentrations in fine-grained sediments near the shelf edge (Table 25-1). The sulfur content of these sediments (0.01-0.13 percent) is quite low compared to other marine sediments (Didyk et al. 1978) and indicates relatively oxidizing conditions within the sediments.

The organic carbon content of samples collected in

1977 (mostly offshore) from Norton Sound is generally lower than those collected in the previous year (Table 25-2). The organic carbon and hydrocarbons are higher in nearshore sediments. Apparently the organic carbon content in this region, unlike that of the southeastern Bering Sea sediments, is generally related to the distance from the presumed terrigenous source, the Yukon River.

The total hydrocarbon/organic carbon (HC/OC) ratio of these samples varies between 0.0002 and 0.005 (Tables 25-1 and 25-2) and is in the range reported for other unpolluted sediments (Palacas et al. 1976). The only exception is EBBS 35 which has an HC/OC ratio of 0.059, similar to those obtained in the Gulf of Mexico (Gearing et al. 1976) and South-

Figure 25-2. Histograms of *n*-alkanes (-) and C_{19} and C_{20} isoprenoids (---) distributions in selected southeastern Bering Sea surficial sediments.

Figure 25-3. Gas chromatograms of hexane fraction from southeastern Bering Sea (EBBS) and Norton Sound (NS) sediments. Station 823 (contaminated with weathered petroleum) Venkatesan et al. 1980; Station 727 (contaminated with fresh petroleum), Kaplan et al., 1976. Numbers 15-33 refer to carbon-chain length of *n*-alkanes. Pr: pristane. UCM: unresolved complex mixture. I.S.: Internal Standard hexamethyl benzene.

Figure 25-4. Histograms of *n*-alkanes (-) and C_{19} and C_{20} isoprenoids (---) distributions in selected Norton Sound surficial sediments. 1976 stations = 47, 88, 131, 156, and 166; 1977 stations = 41 and 14; 1979 stations = 13 and 15.

Figure 25-5. Gas chromatograms of hexane fractions extracted from eelgrass (*Zostera marina*) leaves and surficial sediments from Izembek Lagoon. Numbers 15-31 refer to carbon-chain length of *n*-alkanes.

TABLE 25-3.

Carbon isotopic analysis of humic acids isolated from eastern Bering Sea and Izembek Lagoon surficial sediments

Sample	$\delta^{13}\text{C}_{\text{PDB}}$	Source
EBBS 17	-21.0‰	Peters (unpublished)
EBBS 24	-21.5‰	Peters (unpublished)
EBBS 35	-21.1‰	Peters (unpublished)
EBBS 65	-21.5‰	Peters (unpublished)
Izembek Lagoon	-18.4‰	Stuermer et al. (1978)
Eelgrass	-10.3‰	McConnaughey and McRoy (1979)

ern California (Venkatesan et al. 1980), suggesting pollution of the sediment. The total *n*-alkanes to organic carbon (*n*-alkanes/OC) ratio is less than 0.0007 for all samples (Tables 25-1 and 25-2). Much higher ratios would be expected if unweathered petroleum were present in the sediments (Palacas et al. 1976). It should be stressed that this criterion is only useful in indicating the presence of *unweathered* petroleum in sediments. The absence of resolvable *n*-alkanes in sample 35 (*Alk*/OC $\cong$ 0) could, however, indicate the presence of *weathered* petroleum, as in sediments in nearshore basins of Southern California reported by Venkatesan et al. (1980).

Gas chromatographic data

Southeastern Bering Sea

The gas chromatographic data indicate that allochthonous lipids are the predominant source of hydrocarbons in shelf sediments (Fig. 25-2). This material may be derived from the Kuskokwim or Nushagak Rivers or possibly the coastal lagoons along the shores of the Alaska Peninsula. The drainage basins are in mixed spruce-alder woodlands, which are possible sources of the high molecular weight hydrocarbons in the shelf sediments. The contribution of organic detritus and associated hydrocarbons from the coastal lagoons will be discussed later. The abundance of allochthonous hydrocarbons is generally related to the amount of total sedimentary organic material rather than distance from the presumed source—the two rivers. This is evident from the correlation of total *n*-alkanes with organic carbon. Both organic carbon and *n*-alkanes occur at their highest concentrations in the fine-grained sediments near the center and edge of the shelf (i.e., stations 54, 64, and 38). It is surprising to find the allochthonous alkanes distributed throughout the shelf sediments, since the surface currents in the eastern Bering Sea sweep Kuskokwim River water north to Norton Sound and Bering Strait. The deposition of allochthonous

alkanes in the shelf sediments may in part be related to processes involving ice formation and spring breakup.

The *n*-alkane distributions of the samples analyzed from the Bering Shelf are quite different from sediments contaminated with weathered or fresh petroleum (Fig. 25-3). The homologous series of isoprenoids is not found in the samples. The fact that pristane is much more abundant than phytane (Pr/Ph ratios range from 2 to 18, Table 25-1) suggests that the isoprenoids are derived from varying amounts of biogenic materials rather than from petroleum. The relatively high concentration of pristane in Station 65 may have been derived from the presence of *Calanus* species, which are rich in pristane (Blumer et al. 1964), in the overlying water column. These subarctic oceanic copepods have been reported to be predominantly distributed in the shelf region, north of the Pribilof Islands (Motoda and Minoda 1974). These samples lack UCM, unlike sediments known to have been contaminated by petroleum (Farrington et al. 1977, Crisp et al. 1979, Venkatesan et al. 1980).

The only exception is sample 35 as presented in Fig. 25-3, whose gas chromatogram is characterized by a broad UCM in the entire elution range, demonstrating no resolved hydrocarbons. This pattern is typical of weathered petroleum contamination and is similar to the sediments from Southern California nearshore basins (Venkatesan et al. 1980), a conclusion consistent with the anomalous HC/OC ratio of this sample. The source of these hydrocarbons may be natural submarine seepage, although none have been reported in the southern Bering Sea shelf. However, faults in this area (Marlow et al. 1976) could allow leakage of petroleum from underlying reservoir rocks. The extensive fishing operations around this area make an anthropogenic origin for these hydrocarbons possible; it seems unlikely, however, since the chromatogram from station 37 (Fig. 25-3) and other samples from this area of

TABLE 25-4
Polycyclic aromatic hydrocarbons in sediment samples analyzed by GC/MS
(ng/g)

	Southeastern Bering Sea		Norton Sound (1976)		Norton Sound (1977)		Norton Sound (1979)
	Station 59	Station 64	Station 131	Station 166	Station 35	Station 43	Station 25
O-Xylene	--	--	--	T	T	--	0.2
Isopropylbenzene	--	--	--	--	--	--	--
n-Propylbenzene	--	--	--	T	T	--	--
Indan	--	--	--	--	--	--	--
1,2,3,4-Tetramethylbenzene	--	--	--	T	--	--	--
Naphthalene	T	T	T	T	T	--	T
2-Methylnaphthalene	T	T	T	T	T	T	T
1-Methylnaphthalene	T	T	T	T	T	T	T
Biphenyl	--	--	T	T	T	T	--
2,6-Dimethylnaphthalene	--	--	T	T	--	--	--
Dimethylnaphthalenes ¹	--	--	T	T	--	--	0.1
Trimethylnaphthalenes	--	--	T	T	T	T	0.3
Fluorene	--	--	T	--	T	T	0.1
Dibenzothiophene	--	--	--	--	--	--	--
Phenanthrene	0.3	1.1	4.2 ²	0.2	1.9 ²	0.8 ²	0.4
Anthracene	--	--	--	--	T	T	--
Methylphenanthrenes	0.5	1.3	0.6	T	--	0.2	0.8
Fluoranthene	--	3.0	T	0.1	0.5	T	T
Pyrene	1.4	5.0	2.2 ³	0.1 ³	0.2	T	0.2
Benz(a)anthracene	--	--	--	--	T	--	--
Chrysene	1.2	0.9	3.0 ²	0.6	T	0.4	1.0
Benz(e)pyrene	--	T	T	T	--	--	T
Benz(a)pyrene	--	--	--	--	T	T	--
Perylene	T	T	9.8	T	T	1.3	9.4
Simonellite	--	--	--	See Pyrene	T	T	T
Cadalene	T	--	0.3	T	--	--	T
Retene	1.8	7.3	2.8 ²	--	T	3.1 ²	1.4

T = trace

¹ Excludes 2,6-dimethylnaphthalene when identified

² Coelutes with unknown compound

³ Coelutes with simonellite

intensive fishing do not show evidence of such petroleum contribution.

The hydrocarbon distribution of samples from stations 12 (Fig. 25-2) and 45B resembles the two-component mixture of sources found in western Gulf of Alaska sediments (Kaplan et al. 1977), namely an allochthonous component represented by the odd C₂₅-C₃₁ *n*-alkanes, a nonselective distribution of *n*-alkanes from C₁₇ to C₂₅ with a narrow UCM maximizing in the C₂₂ region. The homologs < *n*-C₂₄ could be of marine origin and consist of residues from primary production (*n*-C₁₇ and *n*-C₁₉; Han and Calvin 1969) and from microbially-altered algal detritus (Johnson and Calder 1973, Cranwell 1976, Hatcher et al. 1977). These biolipids could also have been derived from the bacteria themselves (Lijmbach 1975). Further, these two stations at the southern edge of the study area are far from the influence of surface currents transporting terrigenous material from the Kuskokwim River. This may explain why these samples have predominant autochthonous alkane distribution.

The predominance of allochthonous biolipids at stations 17, 37, and 38 demonstrates the importance of depositional environment in determining the distribution of hydrocarbons that ultimately accumulate in the sediments. The stations near the edge of the continental shelf and south of the Pribilof Islands are in an area of high biological productivity and would be expected to have a high contribution of marine-derived autochthonous lipids as in Walvis Bay (Boon et al. 1975) and Cariaco Trench (Simoneit 1975). Yet, unlike these anoxic environments, in the oxidizing and high-energy depositional environment of the eastern Bering Sea (Sharma 1974) efficient recycling of autochthonous lipids takes place within the water column or at the sediment-water interface. This degradation and recycling may be the result of physical processes, such as strong currents, waves, or ice-gouging, as well as biological processes, such as burrowing by benthic fauna, which cause extensive reworking of the deposited lipids. The fact that this process apparently leaves the sediment enriched in allochthonous alkanes suggests that these

hydrocarbons could be associated with biologically refractory organic material.

Eelgrass

The dominant biota of the coastal lagoons along the shores of the southeastern Bering Sea is eelgrass, which could contribute nutrients and organic detritus to the shelf (Barsdate et al. 1974, McConnaughey and McRoy 1979). Samples of eelgrass and the sediments within eelgrass lagoons were analyzed to characterize their hydrocarbon distribution for comparison and correlation with those of outer shelf sediments.

Gas chromatograms of hexane fraction from surface (0-10 cm) sediment and eelgrass leaves from Izembek Lagoon are illustrated in Fig. 25-5. The hexane fraction from eelgrass consists almost exclusively of a simple mixture of *n*-alkanes, *n*-C₁₅, *n*-C₁₇, and *n*-C₁₉ and only small amounts of hydrocarbons beyond *n*-C₂₅. The distribution is quite different from that found in most other higher plants (Eglinton and Hamilton 1963); it may be the result of the age of the plants and abundance of heterotrophic bacteria, or it may simply mean that this species of subtidal vascular plant does not produce the cuticular wax. It is interesting that the *n*-alkane distribution is similar to that reported for many species of algae; this may indicate that the alkanes are derived from epiphytic and/or macrophytic algae living on the surface of eelgrass leaves, although they are reported to be seldom above 6 percent of the eelgrass biomass in the main eelgrass beds (McConnaughey and McRoy 1979). Hydrocarbon distribution of the surface sediment within the lagoon (Fig. 25-5) indicates that detritus derived from the decomposition of eelgrass leaves cannot be correlated with the allochthonous hydrocarbon distributions found in the shelf sediments. The chromatogram is characterized by a mixture of *n*-alkanes from C₁₇ to C₃₁, with predominant alkanes *n*-C₂₁ and *n*-C₂₃ and a slight UCM in the *n*-C₂₂ region, which could be a result of microbial decomposition of the eelgrass detritus (Johnson and Calder 1973). This is distinctly different from the hydrocarbon distribution in shelf sediments, which are characterized by a predominance of C₂₅-C₃₁ odd carbon-numbered *n*-alkanes.

Carbon isotopic composition of humic acids and protokerogen isolated from surface sediments of the Bering Sea measured by K. Peters (University of California at Los Angeles, personal communication) and that of humic acids isolated from the surface sediment of Izembek Lagoon by Stuermer et al. (1978) are listed in Table 25-3. The humic acids and protokerogens of surface sediments should have carbon isotopic values characteristic of land plants, since the hydrocarbons appear to be predominantly

terrigenous (assuming the same source for all of them). However, the shelf sediment humic acids have carbon isotopic values from -21 to -22‰ (Table 25-3), in between those expected for marine and terrigenous organic carbon (Nissenbaum and Kaplan 1972, Gearing et al. 1977, Craig 1953).

Humic acid from Izembek Lagoon has a carbon isotopic value of -18‰. This is isotopically lighter than the values of -10‰ reported by Smith and Epstein (1970) for *Zostera marina* (whole plant) from Point Mugu Lagoon, California, and -10.3‰ reported by McConnaughey and McRoy (1979) for eelgrass from Izembek Lagoon; it probably indicates a large contribution of planktonic algal and bacterial carbon to the sediment. Indeed, the $\delta^{13}\text{C}$ values of lagoon animals including zooplankton, infauna, and nekton range from -12‰ to -23.1‰ heavier than the average $\delta^{13}\text{C}$ value for Bering Sea phytoplankton (-24.4‰: McConnaughey and McRoy 1979). These data thus indicate that eelgrass and microalgal carbon may be selectively consumed by a wide variety of organisms. Moreover, the particulate organic carbon (POC) in the lagoon ranges from $\delta = -19.9$ to -12.6 ‰, and POC collected at the mouth of the lagoon from -22.1 to -17.3‰. Thus, the isotopic composition of humic acids from the lagoon appears to represent a mixed contribution of eelgrass, plankton, and epiphyte detritus to the lagoonal sediments. Therefore, the values for $\delta^{13}\text{C}$ of organic carbon between -21 and -22‰ measured in the shelf sediments could also be the result of mixing a terrigenous component and a component from the lagoon containing eelgrass detritus. Thus, although hydrocarbons characteristic of the eelgrass lagoons are not found in the eastern Bering Sea shelf sediments, the more resistant humic substances and protokerogens derived from the eelgrass detritus could be transported to the shelf environment.

Norton Sound

A representative gas chromatogram and histograms of alkane distribution given in Figs. 25-3 and 4 illustrate the predominance of 23, 25, 27, 29, and 31 carbon *n*-alkanes with a maximum at *n*-C₂₇, indicating that land plants contribute most of the organic matter (Table 25-2). Stations 49, 131, and 137, in the sound and closer to shore, have the highest content of lipids and *n*-alkanes (Table 25-2). The lipid and *n*-alkane contents progressively decrease along the northeast to southwest transect (i.e., from stations 131 to 147, Fig. 25-1b). The southwest stations 154 and 156, near the mouth of the Yukon River, are richer in organic content and *n*-alkanes. These may have been derived from terrigenous silt resuspended from the Yukon prodelta which extends

across the mouth of the sound. Hydrocarbons from marine plankton in this region are low. Further south, around stations 162 to 168 and west, offshore in stations 39 to 44 and 47A, the terrigenous detritus is diluted by open-ocean sedimentation. The *n*-alkane content of the 1977 samples is also lower than that of 1976 and 1979 samples as expected because of the increased distance of the former stations from the shore.

Station 49, near Pastol Bay, is richer in lipid content and has nearly four times the *n*-alkane content of station 48, even though these stations are equidistant from the shore. This might be explained by the fact that because station 48 is in a higher-energy region than station 49, dispersion of organic matter into the open ocean may occur more readily. The northern region of Norton Sound seems to be poorer in lipids and also in *n*-alkanes (Stations 70, 88, 105), possibly because there are no major sediment sources near the area and the longshore current movement and sediment transport are directed north toward the Bering Strait.

Pristane and phytane are found in trace amounts at a few stations. The very low quantities detected indicate that zooplankton species like *Calanus* copepods, which contribute substantial pristane to a few stations north of the Pribilof Islands in the southeastern Bering Sea, may be impoverished in this area.

Nine samples from stations 47, 172, 174 (1976), 14-17 (1977), and 18 and 22 (1979), collected near the location where petroleum seepage was suspected (Cline and Holmes 1979) show alkane distribution patterns (Fig. 25-4) uncharacteristic of petroleum.

Gas chromatographic/mass spectrometric data

In stations 8, 41, 59, and 65 from the southeastern Bering Sea, the compounds eluting between the normal alkanes from C_{21} to C_{27} have been identified as C_{21} polyolefins and C_{23} - C_{27} mono- and diolefins. These compounds probably represent a contribution from marine organisms. Branched olefins have also been identified in a few samples. Stations 25, 43, 131, and 169 in Norton Sound contain alkenes such as $C_{18}H_{36}$ and a cyclic alkene, $C_{25}H_{46}$. The predominant alkene at station 169 is a $C_{17}H_{34}$ alkene eluting between *n*- C_{16} and *n*- C_{17} ; pristene is also found in significant amounts. In general, stations from offshore contain more alkenes than those nearshore, indicating that these alkenes are of marine origin.

Retene (I, Appendix) and simonellite (II) are present at levels below 3 ppb in all these stations, the former more abundantly than the latter. These diterpenoids are molecular markers derived from terrigenous resinous plants (Simoneit 1977). Cada-

lene (III), a sesquiterpenoid residue found in minor amounts in the sediments at these stations, can be of mixed marine and terrigenous origin. This compound has also been identified in petroleum (Bendoraitis 1974), and in shales and siltstones from the North Atlantic Ocean (Simoneit and Mazurek 1978a, b).

Multiple homologous series of cycloalkanes (e.g., diterpanes, steranes, and triterpanes) could be of biogenic or petrogenic origin. The extended diterpanes and most triterpanes exhibit a strong fragment ion (base peak) at m/z 191 in their mass spectra. Examples from the eastern Bering Sea and an example from Southern California Bight are shown in Fig. 25-6. The inferred skeleton for the homologs is structure IV, based on the structure proposed for the m/z 191 fragment ion of the series (Anders and Robinson 1971). The tricyclic diterpanes range from $C_{19}H_{34}$ to $C_{26}H_{48}$ in stations EBBS 35, NS 43, 131, and 169 and also in EBBS 65 (Simoneit and Kaplan 1980). In a few other stations, such as EBBS 8 and NS 17, diterpenes are more dominant than diterpanes, where only one or two homologs were detected.

In most of the stations, except EBBS 35, the triterpenoids appear to be derived mainly from bacteria or algae (DeRosa et al. 1971, Cardoso et al. 1976), and consist of $17\beta(H)$ -hop-22(29)-ene (diploptene, V), hop-17(21)-ene (VI), $17\beta(H)$ -22,29,30-trisnorhopane (VII, $R=H$, 17β), and $17\beta(H)$ -hopane (VII, $R=C_3H_7$) and the series of extended $17\beta(H)$ -hopanes ranging from C_{31} to C_{33} , with only minor amounts of the $17\alpha(H)$ -hopanes (VIII). The Alaskan sediments contain predominantly C_{27} , C_{30} , and C_{31} 17β , 21β triterpanes. Several C_{30} triterpenes with a double bond in addition to diploptene have also been detected. Most of the C_{29} triterpanes found are not hopanes and their identity has not yet been determined. A $C_{30}H_{50}$ triterpene with base peak 69 and a molecular ion 410 a.m.u. (not a hopene) is present at most of the stations and is also as abundant as *n*- C_{30} . The reason for its occurrence is not known. The extended hopanes ($>C_{31}$) are present as single C-22 diastereomers. The presence of predominantly $17\beta(H)$ stereomers and of the triterpenes which are present in living organisms indicates that these compounds are of recent biogenic origin. The presence of only small quantities of $17\alpha(H)$ stereomers suggests that there is no input from petroleum components (Dastillung and Albrecht 1976, Simoneit and Kaplan 1980, Venkatesan et al. 1980). For purposes of comparison, an example of Recent sediment from San Pedro Basin, Southern California Bight, is given in Fig. 25-6, where the triterpanes are predominantly the $17\alpha(H)$ homologs, and the ex-

Figure 25-6. Relative distribution histograms of diterpenoids and triterpenoids based upon m/z 191 mass chromatograms. EBBS = southeastern Bering Sea; NS = Norton Sound; San Pedro Basin = 20-25 mm core, Venkatesan et al. 1980. Diastereomers indicated by dotted and continuous lines. I is 17 α (H), 18 α (H), 21 β (H)-28, 30-bisnorhopane.

tended hopanes (VIII) are present as 1:1 mixtures of the C-22 diastereomers (Venkatesan et al. 1980). There, the dominant homolog is 17 α (H), 18 β (H), 21 β (H)-28, 30-bisnorhopane ($C_{28}H_{48}$) which has been proposed to be a molecular marker of Southern California petroleum (Seifert et al. 1978, Simoneit and Kaplan 1980). Stations from the eastern Bering Sea contain no C_{28} triterpenoid. The triterpenoidal distribution of station NS 17 (Fig. 25-6) clearly indicates that it is not contaminated with petroleum,

although it is believed to be near a suspected petroleum seep. The only exception is station EBBS 35 (Fig. 25-6), which shows a triterpenoidal distribution very similar to petroleum-contaminated Southern California sediments. This is consistent with the observed n -alkane distribution pattern of this station, typical of weathered petroleum. This sample consists predominantly of 17 α (H) homologs and the extended hopanes are present as 1:1 mixtures of the C-22 diastereomers. The C_{28} bisnorhopane found in

Southern California petroleum is also found in this sample, but is much less abundant than in Southern California Bight sediments (Venkatesan et al. 1980).

The resolved polynuclear aromatic compounds (PAH) are $\leq 1.0 \mu\text{g/g}$ sediment in these stations, much less than those found in Beaufort Sea sediments (Kaplan et al. 1979). Concentrations of selected PAH compounds are presented in Table 25-4. Generally, biphenyl, naphthalenes, phenanthrene, fluorene, fluoranthene, pyrene, chrysene, and traces of their alkyl substituted homologs have been detected by GC/MS analyses. Benzo(E)pyrene has also been found in all the samples. In general, the parent PAH compounds are more abundant than their alkylated homologs (Table 25-4), indicating that they come from pyrolytic sources rather than from crude oil shales (Coleman et al. 1973, Youngblood and Blumer 1975). A more detailed study is required before any positive conclusions can be reached on sources and mechanisms of distribution.

Perylene is more abundant (~ 10 ppb in station 131) in nearshore sediments than in offshore sediments (traces in stations 35 and 166). Perylene could be generated *in situ* by transformation of some terrestrial precursor compound (Bergmann et al. 1964). Quinones transported from soils and originating from plants could form perylene (Aizenshtat 1973) in reducing micropaleoenvironmentary conditions as described by Welte and Ebhardt (1968), even though the overall sediment is oxidizing. Coronene and 1,12 benzoperylene have also been detected in these samples.

CONCLUSIONS

The alkanes in sediments of the study area generally show a biomodal distribution typical of a mixture of allochthonous and autochthonous sources. In short, the absence of unresolved complex mixture and the distribution pattern of *n*-alkanes and extended triterpanes in most of the stations studied can be attributed to recent biogenic sources characteristic of unpolluted environments.

The allochthonous lipids, the primary source of hydrocarbons in the surface sediments, are probably transported to the continental shelf by river discharge and erosion and redistribution of surface sediments. Correlation of the hydrocarbon distribution of the eastern Bering Sea with the hydrocarbons extracted from eelgrass (*Zostera marina*) and sediments from within Izembek Lagoon indicates that the latter environment may not be a significant source of hydrocarbons in the outer shelf sediments. However, carbon isotopic analysis of humic and kerogenous substances from the lagoon and shelf sediments

indicates that these biologically refractory organic materials may be transported to the shelf environment.

The presence of relatively small amounts of autochthonous hydrocarbons in the sediments, in spite of the high biological productivity of the region (especially the southeastern Bering Sea), suggests rapid and efficient recycling of marine lipids within the water column or at the sediment-water interface. Presence of higher concentrations of relatively labile hydrocarbons derived from the autochthonous sources, identified in only a few stations on the Bering Sea shelf, may be important in an assessment of the fate and effects of petroleum products introduced into this marine environment. Any petroleum contamination and deposition in those environments may last a long time.

The stations in Norton Sound suspected to be near natural gas seeps do not show hydrocarbon distribution patterns characteristic of petroleum.

The distribution of PAH compounds is more complex than that of *n*-alkanes and appears to be predominantly pyrolytic, possibly derived from forest fires.

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ADDENDUM: Recently Kvenvolden et al. (1979) have reported the seep near Station 17, Norton Sound, to be thermogenic in origin with CO_2 as its major component.

APPENDIX: STRUCTURES CITED IN THE TEXT

I. retene, $C_{18}H_{18}$

II. simonellite, $C_{19}H_{24}$

III. cadalene, $C_{15}H_{18}$

IV. extended diterpanes
 $R = C_2H_5 - C_{12}H_{25}$

V. diploptene, $C_{30}H_{50}$

VI. hop-17(21)-ene, $C_{30}H_{50}$

VII. $17\beta(H)$,
 $21\beta(H)$ -hopanes
 $R = H, C_2H_5, C_3H_7$

VIII. extended $17\alpha(H)$,
 $21\beta(H)$ -hopanes
 $R = CH_3 - C_5H_{11}$

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Hydrocarbon Gases in Near-surface Sediment of the Northern Bering Sea

26

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ABSTRACT

Methane, ethane, ethene, propane, propene, *n*-butane, and isobutane are common in bottom sediment of the northern Bering Sea. At eight sites the content of methane rapidly increases with depth within the first four meters of sediment. These concentration gradients and absolute methane concentrations indicate that the interstitial water of the near-surface sediment at these sites may be gas saturated. These gas-charged sediments may be unstable, creating potential geologic hazards and, in certain areas, causing the formation of seafloor craters.

The isotopic compositions of methane at four of the sites range from -69 to -80 ‰ ($\delta^{13}\text{C}_{\text{PDB}}$). This range of values clearly indicates that the methane derives from microbial processes, possibly within the near-surface Pleistocene peat deposits that are common throughout the northern Bering Sea. At one site in Norton Sound, near-surface sediment is charged with CO_2 , accompanied by minor concentrations of hydrocarbons, that is seeping from the seafloor. Methane in this gas mixture has an isotopic composition of -36 ‰, a value that suggests derivation from thermal processes at depth in Norton Basin.

The presence of sediment charged with methane or CO_2 cannot in general be predicted from analyses of surface sediment, which usually contains hydrocarbon gases and CO_2 at low concentrations. Sampling beneath a sediment depth of about 0.5 m is generally required to detect high concentrations of gas. Acoustic anomalies detected on high-resolution seismic records indicate the presence of gas-charged sediment, but gas analyses of sediment samples from areas with these anomalies do not always confirm that high concentrations of gas are there. Conversely, high concentrations of methane are sometimes found at sites where no acoustic anomalies are obvious on high-resolution records.

INTRODUCTION

About twenty years ago Emery and Hoggan (1958) described the occurrence of hydrocarbon gases in near-surface marine sediment from Santa Barbara Basin, off southern California. These anoxic sediments contain methane, ethane, propane, butanes,

pentanes, and hexanes, with methane being one to almost five orders of magnitude greater in concentration than any of the other hydrocarbons. Geochemical studies that followed have generally focused on methane and the processes that can account for its occurrence and distribution in a variety of aquatic sediments (Reeburgh 1969, Whelan 1974, Martens and Berner 1974, Claypool and Kaplan 1974, Orem-land 1975, Barnes and Goldberg 1976, and Kosiur and Warford 1979). Recently Bernard et al. (1978) described the distribution of methane, ethene, propane, and propene in shelf and slope sediment in the Gulf of Mexico, and Kvenvolden and Redden (1980) reported on the occurrence of these gases in sediment from the outer shelf, slope, and basin of the Bering Sea. The present study examines inner shelf areas of the Bering Sea and considers the hydrocarbon gases methane (C_1), ethane (C_2), ethene ($\text{C}_{2:1}$), propane (C_3), propene ($\text{C}_{3:1}$), isobutane (*i*- C_4), and *n*-butane (*n*- C_4) in sediment of the inner Bering Shelf in Norton Sound and the adjacent eastern Chirikov Basin (Fig. 26-1).

Norton Sound is an elongate, east-west trending bay in the western coast of Alaska bounded on the north by the Seward Peninsula, on the east by the Alaskan mainland, on the south by the Yukon Delta, and on the west by the Chirikov Basin. The floor of the sound is very flat, and the water averages about 20 m deep. To the west in the Chirikov Basin water depths increase to about 50 m, especially in the northern part of the basin at the Bering Strait.

When sea level fell in late Pleistocene time, the floor of Norton Sound and the eastern Chirikov Basin became exposed (Nelson and Hopkins 1972). During

Figure 26-1. Location of hydrocarbon gas sampling sites in Norton Sound and Chirikov Basin (dots). Sites are designated with the last digit of the year when the site was occupied followed by the station number. In parentheses is the interval in centimeters from which samples were taken. After the colon is the number of samples analyzed for hydrocarbon gases in that interval.

this time fluvial processes and tundra vegetation characterized the area (Hopkins 1967), and peaty mud was deposited over much of the region. This mud contains 2-8 percent organic carbon. As sea level rose during latest Pleistocene time, marine sedimentation resumed. In Holocene time, fine-grained, sandy silt derived mainly from the Yukon River blanketed the area with a cover up to 10 m thick (McManus et al. 1977, Nelson and Creager 1977). In contrast to the nonmarine sediment, the organic content of the overlying sediment ranges only from 0.5 to 1.0 percent (Nelson 1977).

METHODS

The procedures used for this work consisted of analyses of sediment samples for hydrocarbon gases and carbon dioxide on board ship and measurements of the carbon isotopic compositions of organic matter, C_1 , and CO_2 in shore-based laboratories (Nelson et al. 1978, Kvenvolden et al. 1979a, b). Sediment samples were recovered by vibracores and surface grab-samplers during three summer field seasons in 1976, 1977, and 1978. Gases were extracted from sediments in the following manner: volumes (about 0.5 l) of wet sediment were extruded into 0.95 l double-friction-seal cans that had two septum-covered holes near the top. Helium-purged distilled water was added to each can until a 100-ml headspace remained. Each can was closed with a lid, and the headspace was purged with helium through the septa. The cans were vigorously shaken for ten minutes to release gases into the headspace. Exactly one millimeter of the headspace gas mixture was analyzed by gas chromatography using both flame ionization (for hydrocarbons) and thermal conductivity (for C_1 at high concentrations and for CO_2) detectors. Concentrations of gases were determined by measurements of peak heights on chromatograms and with quantitative standards. Partition coefficients were used to correct for the different solubilities of the hydrocarbon gases. Concentrations are reported in nl, μ l, or ml per liter of wet sediment. This method of extraction yields semiquantitative results, but, because all samples were processed in the same manner, the results can be compared.

The carbon isotopic composition of organic matter, C_1 , and CO_2 were determined by mass spectrometry. Before mass spectrometry C_1 and CO_2 were separated in a vacuum line-combustion apparatus modified after Craig (1953). Results are reported in parts per mil ($^{\circ}/_{\infty}$) as:

$$\delta^{13}C = \frac{^{13}C/^{12}C \text{ sample} - ^{13}C/^{12}C \text{ standard}}{^{13}C/^{12}C \text{ standard}} \times 1000$$

where the standard is Pee Dee belemnite (PDB). There were insufficient concentrations for carbon isotopic determinations of C_2 and higher hydrocarbons.

RESULTS

C_1 is the most abundant hydrocarbon gas found in the first five meters of sediment in Norton Sound and the eastern Chirikov Basin. Fig. 26-2 shows the geographic distribution of maximum concentrations of C_1 . At eight sites this concentration exceeds 1 ml/l, and at five of these sites (8-4, 8-8, 8-15, 8-21, and 8-22) concentrations exceed 10 ml/l. At the other stations the maximum amount of C_1 measured was less than 100 μ l/l with two exceptions, at 7-33 and 8-6, where it was about 200 μ l/l at each site. The amount of C_1 found depends to some extent on the depth of core, for concentrations of C_1 generally increase with depth. Thus, surface samples and short cores usually have lower amounts of C_1 than do samples taken from greater depths. The vertical distribution of C_1 at the eight sites mentioned above is shown in Fig. 26-3. For these sites the concentration of C_1 increases by four or five orders of magnitude within the top four or five meters of sediment. Also shown is the distribution of methane at sites 7-17 and 8-3. The data for these sites are combined because the sites are at essentially the same location sampled during two different field seasons. Here the C_1 concentrations, especially in deeper samples, are much lower than at the eight sites. The major gas at 7-17/8-3 is not C_1 but CO_2 .

The second most abundant hydrocarbon gas in this area is C_2 ; C_3 concentrations are usually slightly lower and generally parallel C_2 concentration profiles with depth. The geographic distribution of $C_2 + C_3$ is shown in Fig. 26-4. The maximum concentrations of these two hydrocarbons are usually less than 1 μ l/l. At two sites, 8-4 and 8-17, maximum $C_2 + C_3$ concentrations are slightly higher (1.2 and 1.1 μ l/l, respectively), but at site 7-17/8-3, $C_2 + C_3$ maximum concentration is almost 8 μ l/l. At the eight sites where C_1 shows 4-5 orders of magnitude increase in concentrations with depth, $C_2 + C_3$ increased by about two orders, but the profiles of concentration are more variable (Fig. 26-5) than the C_1 concentration profile (Fig. 26-3). The concentrations of $C_2 + C_3$ in samples from 7-17/8-3 are much higher than in all other samples.

Both $C_{2:1}$ and $C_{3:1}$ are present in all samples analyzed. Concentrations are variable, but as a general rule $C_{2:1}$ exceeds C_2 in surface samples and with increasing depth the reverse is observed. At site 7-17/8-3, C_2 is much more abundant than $C_{2:1}$, with

Figure 26-2. Distribution of maximum concentrations of C_1 in $\mu\text{L/l}$ and mL/l of wet sediment at each site.

Figure 26-3. Graph of concentrations of C_1 in $\mu\text{L/l}$ and mL/l of wet sediment vs. depth (cm) for sediment samples from cores taken at nine sites in Norton Sound and eastern Chirikov Basin. Sites 7-17 and 8-3 are the same location; results are combined into one curve.

Figure 26-4. Distribution of maximum concentrations of $C_2 + C_3$ in nL/l and μ L/l of wet sediment at each site.

Figure 26-5. Graph of concentrations of $C_2 + C_3$ in nL/l and μ l/l of wet sediment vs. depth (cm) for sediment samples from cores taken at nine sites in Norton Sound and eastern Chirikov Basin.

the $C_2/C_{2:1}$ ratio reaching a maximum of 340 at a depth of 60 cm. A similar relation holds for $C_{3:1}$ and C_3 . At the surface $C_{3:1}$ is usually more abundant than C_3 , and the reverse is true for deeper samples. At 7-17/8-3, C_3 is always more abundant than $C_{3:1}$ —by a factor of 47 at a depth of 200 cm.

The concentrations of $i-C_4$ and $n-C_4$ are lower than the lighter hydrocarbon gases and in many cases reach the limit of detection of the method, about 2 nl/l. In general the concentrations of $i-C_4 + n-C_4$ are less than 100 nl/l, and the distribution with depth is variable. In samples from 7-17/8-3, concentrations of $i-C_4 + n-C_4$ reach a maximum of 14 μ l/l at a depth of 200 cm.

Hydrocarbons from C_5 to C_7 are measured in a single backflush peak from chromatography and are designated C_{5+} . C_{5+} hydrocarbons commonly occur in high concentrations in surface samples from this area and particularly in core samples from 7-17/8-3.

BIOGENIC METHANE

The occurrence in anoxic sediment of high concentrations of C_1 resulting from microbial decomposition of organic matter is well established (Emery and Hoggan 1958, Barnes and Goldberg 1976, Reeburgh and Heggie 1977, and Kosiur and Warford 1979). C_1 is both produced and consumed by microorganisms, and models for these processes have been devised (Claypool and Kaplan 1974, Martens and Berner 1974, Barnes and Goldberg 1976, Kosiur and Warford 1979). In less reducing sediments of open marine environments, C_1 is also present but at concentrations as much as five orders of magnitude less than those observed in anoxic marine sediments (Bernard et al. 1978, Kvenvolden and Redden 1980). Although there is much less C_1 in sediments of open marine environments, the processes that generate gas are probably similar to those in anoxic sediments, but much slower.

At eight sites in Norton Sound and the eastern Chirikov Basin abundances of C_1 increase by four or five orders of magnitude within the first five meters of sediment, reaching concentrations near or exceeding saturation of the interstitial water. These shallow sediments are probably anoxic, and the C_1 probably is being generated by the decomposition of peaty mud that contains 2-8 percent organic carbon, and is buried under marine sediment of lower carbon content (0.5-1.0 percent). This sediment cover is thickest near the front of the Yukon Delta and thins to the north (McManus et al. 1977, Nelson and Creager 1977, Nelson 1977). The depth of burial of

the peaty mud may account for the groupings of the C_1 concentration profiles shown in Fig. 26-3. Seven of the sites (6-121, 6-125, 6-131, 8-4, 8-8, 8-15, and 8-21) have profiles that group together. These seven sites are located in the eastern and northern parts of Norton Sound and in the Chirikov Basin near Port Clarence (Fig. 26-2). In these areas, peaty sediment is buried under about 2 m of sandy silt. The seven profiles show maximum concentrations below about 1.5 m. Therefore, if peaty mud is the source of the methane, the depth of its burial accounts for the depth at which high C_1 concentrations are found. In contrast, one C_1 concentration profile (8-22) reaches maximum values below 3 m. This site was northwest of the Yukon Delta in the southern part of Norton Sound, where the sediment cover is thicker and the peaty mud is more deeply buried. Thus there is a correlation between the depth of buried organic matter and the depth at which C_1 concentrations reach high values. At site 7-17/8-3, C_1 concentrations follow a different trend to be discussed below.

That the high concentrations of C_1 at eight sites result from microbiological processes is supported by both chemical and isotopic data. Higher molecular weight hydrocarbons accompany C_1 , and the ratio $C_1/(C_2 + C_3)$ can be used as a guide to interpret mode of formation. Likewise, the carbon isotopic composition of C_1 can be used to interpret process of formation (Bernard et al. 1976, 1977). Microbial degradation of organic matter produces hydrocarbon gases with $C_1/(C_2 + C_3)$ ratios greater than 1,000, and with $\delta^{13}C_{1PDB}$ lighter than -60‰ . Table 26-1 shows these parameters for samples from the eight sites. On the basis of the criteria stated above, it is clear that the C_1 at the eight sites was derived from microbiological processes, and the buried peaty mud, in which the organic carbon has an isotopic composition of -28‰ (Kvenvolden et al. 1979a, b), is the likely source.

Other sites may exist in Norton Sound and eastern Chirikov Basin where C_1 concentrations exceed 1 ml/l. Finding these locations will require sampling below about 1 m of sediment (3 m or more off the Yukon Delta), because the occurrence of high amounts of C_1 at shallow depths is not manifest at the surface. Either the surface layer of sandy silt seals the C_1 , preventing its migration to the surface, or the rate of consumption or diffusion of C_1 in the upper meter is very rapid, leading to low concentrations of C_1 at the surface. At two sites, 8-6 and 7-33 (Fig. 26-2), maximum concentrations of C_1 of 224 and 196 μ l/l respectively, may hint that much higher concentrations are present at greater depths. At site 7-33, the deepest sample (Fig. 26-1) came from 70

TABLE 26-1

$C_1/(C_2+C_3)$ ratios and $\delta^{13}C_1$ values for samples containing C_1 concentrations in excess of 1 ml/l

Site	Maximum $C_1/(C_2+C_3)$	$\delta^{13}C_1$ * (‰)
8-4	24000	-80 ¹
8-8	71000	nd
8-15	28000	nd
8-21	440000	nd
8-22	88000	nd
6-121	6500	-72 ²
6-125	28000	-69 ²
6-131	5400	-75 ²

*relative to the PDB standard

¹ Kvenvolden et al. (1979a, b)

² Nelson et al. (1979)

cm. If this sample, containing 196 μ l/l, were plotted on Fig. 26-3, it would fall within the envelope of C_1 -concentration profiles of cores in which C_1 exceeds 1 ml/l at depth. The case for site 8-6 is not so clear. The sample containing 224 μ l/l comes from a depth of 220 cm (Fig. 26-1). If plotted on Fig. 26-3, this value would fall below the envelope of C_1 -concentration profiles. At site 8-6, peaty mud may be more deeply buried than at other sites in northern Norton Sound. Only deeper sampling can directly verify the presence of higher amounts of C_1 . At other sites in Norton Sound and eastern Chirikov Basin, C_1 concentrations are below 100 μ l/l and at many sites below 10 μ l/l (Fig. 26-2).

POSSIBLE BIOLOGIC ORIGIN OF OTHER HYDROCARBONS

Besides C_1 , other hydrocarbon gases are present in these sediments, but in much lower quantities than C_1 . The maximum concentrations of $C_2 + C_3$ are in the same range as the minimum concentrations of C_1 . At sites where C_1 increases rapidly with depth (Fig. 26-3), $C_2 + C_3$ also generally increases (Fig. 26-5), but much more slowly than C_1 . Concentrations of $i-C_4$ and $n-C_4$ are even lower than concentrations of $C_2 + C_3$, and the $i-C_4 + n-C_4$ concentrations are erratic with depth. As a generalization, however, the abundances of the higher hydrocarbons, C_2 , C_3 , and C_4 , are greater in samples where concentrations of C_1 are greater. Therefore, the processes that produce C_1 may also be responsible in part for the generation of the higher hydrocarbons. Microbiological production and consumption provides a reasonable mechanism to

account for C_1 at the eight sites where C_1 concentrations increase beyond 1 ml/l. Therefore, microbial processes may also explain the occurrence of the higher molecular weight hydrocarbons, although evidence for this process remains circumstantial. Laboratory experiments have demonstrated the microbial formation of C_2 and C_3 (Davis and Squires 1954). Thus there is support for the suggestion that the C_2 and the C_3 hydrocarbons at these sites can come from microbial processes, but there is no precedent in the microbiological literature for the production of C_4 hydrocarbons.

The presence and distribution of $C_{2:1}$ are probably controlled by biological processes, but these processes likely differ from those which account for the very high C_1 concentrations. These unsaturated hydrocarbons have been formed by microbial action in the laboratory (Davis and Squires 1954), and $C_{2:1}$ is produced in soils by bacteria (Primrose and Dilworth 1976). In the sediment the process seems to take place uniformly, because there is no obvious concentration gradient with depth. In surface samples concentrations of $C_{2:1}$ and $C_{3:1}$ are higher respectively than concentrations of C_2 and C_3 . With depth, concentrations of C_2 and C_3 increase slightly so that below the surface, ratios of $C_2/C_{2:1}$ and $C_3/C_{3:1}$ are usually equal to or greater than one.

THERMOGENIC HYDROCARBONS

The above discussion focused mainly on sites where C_1 concentration increases rapidly with depth and consideration has been given to the heavier hydrocarbons associated with this C_1 . At one site, 7-17/8-3, however, C_1 concentrations are not unusually high—less than 100 μ l/l (Fig. 26-3)—but concentrations of $C_2 + C_3$ (Fig. 26-5) and $i-C_4 + n-C_4$ are unusually large relative to concentrations seen elsewhere in the sediments of this area, or for that matter, anywhere else in marine sediments off Alaska.

Site 7-17/8-3 has been studied in great detail since anomalous hydrocarbon concentrations were first discovered in the water column at the site (Cline and Holmes 1977). Nelson et al. (1978) showed that the sediments here also contain anomalous hydrocarbon concentrations. Kvenvolden et al. (1979a, b) confirmed the hydrocarbon chemistry and discovered that the major gas component within the sediment and escaping into the water column is CO_2 . The $C_1/(C_2 + C_3)$ ratios in sediment at this site are less than 10 and the $\delta^{13}C_1$ is -36‰. These numbers differ greatly from those discussed earlier where microbiological processes were inferred.

The hydrocarbons at site 7-17/8-3 are probably

derived from thermochemical processes, judging from the molecular distribution of C_1 , C_2 , and C_3 and the isotopic composition of C_1 . In addition, anomalously high concentrations of $i-C_4$, $n-C_4$, and C_{5+} (gasoline-range hydrocarbons) support the mechanism of thermochemical processes (Kvenvolden and Claypool 1980). Heat for this process must be available at depth in Norton Basin. The hydrocarbons resulting from the thermal decomposition of organic matter within the basin must migrate along with CO_2 up fault zones to the surface and escape as a seep. Although the hydrocarbon chemistry indicates that the hydrocarbons at site 7-17/8-3 probably migrate from depth, the concentration profiles (Figs. 26-3 and 26-5) indicate that special conditions of migration must exist. The fact that hydrocarbons are leaking into the water column suggests that surface sediments should contain large amounts of these hydrocarbons. On the contrary, surface samples at this site contain very low concentrations of hydrocarbons. In fact, surface samples (0-10 cm) at this site show no evidence of the high concentrations of hydrocarbons deeper in the sediment. The gradients of C_1 (Fig. 26-3), $C_2 + C_3$ (Fig. 26-5), and $i-C_4 + n-C_4$ decrease rapidly toward the sediment surface. This rapid decrease and lack of significant quantities of hydrocarbons at the sediment surface can be explained either by rapid diffusion of hydrocarbons into the water column from the first few centimeters of sediment or by the presence of discrete gas vents that pipe the hydrocarbons through the sediment, leaving few hydrocarbons remaining in the sediment. The second explanation is more reasonable, because active gas vents were seen by television in 1978 (Kvenvolden et al. 1979a). The concentration profiles (Figs. 26-3 and 26-5) at site 7-17/8-3 reach a maximum value at a depth of about 1-2 m and then decrease. This profile suggests that migration from greater depths does not involve diffusion of hydrocarbons within the underlying sediment but rather that the hydrocarbons are following distinct conduits such as faults. Near the surface the hydrocarbons are dispersed into the sediment where they eventually vent along with CO_2 into the water. That the hydrocarbons from this seep are present in the water column has been documented by Cline and Holmes (1977). The waters of Norton Sound and the eastern Chirikov Basin also contain a regional distribution of hydrocarbon gases (Cline et al. 1978), the sources of which, in part, may be the underlying surface and near-surface sediment.

GEOPHYSICAL EVIDENCE

The presence of gas in near-surface sediments can

cause acoustic anomalies on high-resolution geophysical records where the gas is no longer in solution in the interstitial water but takes the form of bubbles. Schubel (1974), for example, demonstrated how high concentrations of gas affect acoustic properties of sediments. At the eight sites where C_1 concentrations exceeded 1 ml/l and may have reached and exceeded interstitial water solubility, bubble-phase C_1 may be present. Acoustic anomalies would be expected on high-resolution records from these sites if free gas is indeed present.

Geophysical transects, using 800-J boomer, 3.5 kHz subbottom profiler, and 120 kJ sparker systems, indicate that near-surface acoustic anomalies are widespread in Norton Sound. Fig. 26-6 shows those sites, sampled for hydrocarbon gases, at which acoustic anomalies are seen on geophysical records. At three sites (6-125, 8-4, and 8-21), acoustic anomalies correspond to samples having high concentrations of C_1 . The acoustic anomaly and associated C_1 at 8-4 were discussed in detail by Kvenvolden et al. (1979 a, b). The characteristics of the acoustic anomaly at 8-21 suggest that the cause may be controlled more by the presence of glacial till deposits than by gas. At sites 7-17/8-3, geophysical records show both near-surface and deeper acoustic anomalies. There the sediment is charged with CO_2 rather than C_1 , and the CO_2 is escaping from the seafloor as a submarine seep, observed acoustically and by television (Kvenvolden et al. 1979a). At five sites (6-121, 6-131, 8-8, 8-15, and 8-22) on Fig. 26-6, where high concentrations of C_1 were measured, no acoustic anomalies were detected. At sites 6-121 and 6-131 no geophysical records were obtained; it is uncertain, therefore, whether or not acoustic anomalies are present at these sites. At sites 8-8, 8-15, and 8-22, high-resolution geophysical records show no evidence of acoustic anomalies, although the geochemical measurements indicate high concentrations of C_1 (Figs. 26-2 and 26-4). Apparently the gas concentrations at these sites were not in the proper range to produce anomalies on the records of the geophysical systems employed. On the other hand, acoustic anomalies were observed on records at sites 8-1, 8-9, and 8-10, but maximum C_1 abundances in cores at these sites were only 6, 4, and 15 μ l/l, respectively. Concentrations this low are not expected to produce acoustic anomalies. In addition, acoustic anomalies were found in geophysical records at sites 6-168, 7-22, and 7-25, but sampling at these sites was not deep enough (Fig. 26-1) to test the presence of high C_1 or CO_2 concentrations at depth.

High concentrations of C_1 in near-surface sediment may cause instability and may lead to crater forma-

Figure 26-6. Distribution of acoustic anomalies and sites with high concentrations of gas.

tion whenever the gas vents abruptly into the water column. Nelson et al. (1978) discussed some preliminary engineering information related to the stability of gas-charged sediment in Norton Sound, and Nelson et al. (1979) proposed that the craters found in central Norton Sound may result from the rapid escape of gas from gas-charged sediments.

SUMMARY

Hydrocarbon gases, methane, ethane, ethene, propane, propene, isobutane, and *n*-butane are common in surface and near-surface sediment of Norton Sound and eastern Chirikov Basin. Quantitatively, methane is the most important hydrocarbon gas. At eight sites in this area methane abundances increase with depth in the sediment by four or five orders of magnitude, reaching concentrations near or exceeding saturation of the interstitial water. The highest value measured is about 55 ml/l. This methane is probably being generated by microbial decomposition of peat that is buried with mud under a cover of fine-grained, sandy silt derived from the Yukon Delta. Maximum ratios of methane to ethane plus propane at the eight sites are large, ranging from about 5×10^3 to 440×10^3 . Carbon isotopic compositions range from -69 to -80‰ (relative to the PDB standard). These molecular and isotopic compositions strongly suggest that microbiological processes are involved in the production of at least methane. Higher molecular weight hydrocarbons are present in much lower concentrations than methane; but, as a general rule over the area, the trends in concentrations of the hydrocarbons, at least through propane, are roughly the same. Therefore, microbiological processes may also be responsible for the hydrocarbon gases heavier than methane.

At one site in Norton Sound the concentrations of all hydrocarbon gases are anomalous. For example, methane is anomalously low in concentration relative to the methane at the eight sites mentioned above. On the other hand, ethane, propane, and the butanes are all anomalously high in concentration. The ratios of methane to ethane plus propane are less than 10 and the isotopic composition of methane is about -36‰ . The magnitude of these parameters sharply contrasts with the values obtained elsewhere in the area and indicates that thermochemical rather than biochemical processes are at work. The maximum concentration of hydrocarbons from samples at this site is about 100 $\mu\text{l/l}$. The major component is carbon dioxide. The carbon dioxide and hydrocarbons are actively seeping from the sediment into the water column.

A number of geological consequences result from the presence of gases in near-surface sediment of this area. Where gas concentrations approach and exceed their solubilities in the interstitial water, free gas in the form of bubbles occurs. This gas modifies the acoustic properties of the sediments, and near-surface acoustic anomalies are detected on high-resolution geophysical records. Acoustic anomalies were observed at three of the eight sites where methane is very abundant and also at the site where carbon dioxide was observed along with hydrocarbons. On the other hand, acoustic anomalies were also seen in geophysical records from three sites where maximum gas concentrations in core samples were too low to cause anomalies. At three other sites there was no geophysical evidence for acoustic anomalies although geochemical measurements indicated high methane concentrations in the sediment. The presence of high gas concentrations in near-surface sediments can lead to sediment instability and the possibility of seafloor cratering. Finally, the thermochemical hydrocarbons observed at one site may have been produced deep within Norton Basin and be migrating to the surface. These hydrocarbons have petroleum-like characteristics and may indicate petroleum generation and accumulation at depth.

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Distribution of Dissolved LMW Hydrocarbons in Bristol Bay, Alaska: Implications for Future Gas and Oil Development

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ABSTRACT

In September and October, 1975, and again in July, 1976, the distribution of dissolved low molecular weight hydrocarbons was determined in Bristol Bay, Alaska. The concentrations were relatively low compared to other Alaskan shelf areas and show a significant seasonal signature. Local production of methane is accelerated in summer as it is for the alkenes. The concentrations of ethane and ethene are in linear relation in summer, suggesting a common source or perhaps a common organic precursor. The distribution of methane is strongly coupled to circulation and, in particular, to the location of hydrographic fronts. In contrast, the alkenes appear to be regulated more by biological activity than circulation. In composition, LMW hydrocarbons arising from a thermogenic source can be readily distinguished from their biological equivalents on the basis of the relative concentrations of ethane and ethene. Elementary modeling of a line hydrocarbon source suggests that hydrocarbon trajectories could be traced for several hundred km, assuming a source concentration 100 times above ambient levels. Other model scenarios are also considered.

INTRODUCTION

The low molecular weight (LMW) alkanes (i.e., methane, ethane, propane, iso- and *n*-butanes) are abundant constituents of crude oil and natural gas (Clark and Brown 1977). As such, they may indicate the presence of crude oil and thermogenic gas in marine waters. These gases may originate in production activities associated with offshore drilling, venting, and transportation and transfer operations (Brooks and Sackett 1973 and 1977, Bernard et al. 1976), or they may arise from natural hydrocarbon seeps (Carlisle et al. 1975, Dunlap et al. 1960, Sackett 1977, Cline and Holmes 1977).

The LMW alkanes represent a significant fraction of many crude oils, and they also are produced in small but significant amounts in marine waters and

sediments (Brooks and Sackett 1973, Swinnerton and Lamontagne 1974, Bernard et al. 1978). Methane, the most abundant LMW hydrocarbon, is produced through the fermentation of simple organic acids or in hydrogen reduction of CO₂ by anaerobic microorganisms (McCarty 1964, Wolfe 1971, Reeburgh and Heggie 1977). Recent evidence suggests that methane is also produced in oxic marine waters, presumably from organisms living in reducing microenvironments (Scranton and Brewer 1977).

On the other hand, the origin of the C₂-C₄ compounds is less clear. It is known, for example, that ethene is produced by soil bacteria (Smith and Cook 1974), but the significance of these organisms in ocean waters is not known. What is known, however, is that the alkenes ethene and propene are usually enriched in surface ocean waters (Swinnerton and Lamontagne 1974, Cline et al. 1978). Production of these compounds in marine surface waters appears to be a photochemical process involving dissolved organic carbon (Wilson et al. 1970), but a biological contribution from microorganisms cannot be completely dismissed on the basis of currently available data (Lamontagne et al. 1975).

The purpose of this chapter is to present the distributions and abundances of dissolved LMW hydrocarbons in Bristol Bay, Alaska. The study was carried out in September and October 1975 and July 1976. Emphasis is on the natural occurrences, their relationships to circulation and source regions, and the potential usefulness of these compounds in tracing both catastrophic and chronic petroleum contamination arising from future resource development. This program was developed in response to objectives set

forth in the Environmental Study Plan for the Gulf of Alaska, Southeastern Bering Sea, and the Beaufort Sea (DOI/DOC 1975).

SAMPLING AND CHROMATOGRAPHIC ANALYSIS

Water was collected in PVC Niskin® samplers attached to a Plessey model 9040 CTD-rosette system deployed at predetermined depths. Upon retrieval, water was carefully transferred to 1-liter glass-stoppered bottles so as not to include bubbles, allowing one volume to overflow for rinsing. To inhibit biological activity, 100-200 mg sodium azide was added to each sample. Samples were stored at ambient temperatures (approximately 5-10 C) in the dark until analyses could be accomplished, usually within two hours of sampling.

Dissolved low molecular weight hydrocarbons, C_1 - C_4 , were quantitatively removed from solution by a modification of a procedure originally proposed by Swinnerton and Linnenbom (1967). In the modified procedure, hydrocarbons were removed in a stream of helium (approximately 100 ml/min) and concentrated on a single Activated Alumina® trap (0.64 cm o.d. $\times$ 5 cm) held at -196 C (LN_2). Quantitative stripping of the hydrocarbons was achieved in 10 minutes at which time the trap was warmed to 100 C and the hydrocarbons injected into a gas chromatograph (Hewlett-Packard model 5711) equipped with dual flame ionization detectors. Chromatographic separation of the components was originally effected on a Poropak® Q column (0.48 cm o.d. $\times$ 2.5 m) held isothermally at 30 C. With a helium carrier flow rate of 60 ml/min, analysis was completed in 15 minutes. During the second cruise (July 1976), the chromatographic column was modified by adding a short activated alumina column (0.48 cm o.d. $\times$ 5 cm) impregnated with 1 percent silver nitrate by weight. This modification, coupled with temperature programming from 110 C to 150 C, resulted in sharper peaks, improved separation of the alkenes, and shorter analysis time (Cline and Feely 1976).

Calibration was carried out by injecting and trapping 1-ml volumes of a certified standard hydrocarbon mixture prepared by the Matheson Co. This standard mixture was subsequently recalibrated by NBS and found to conform to the previously stated accuracy of 10 percent for each component.

Replicate analyses of water samples were carried out at three stations in Bristol Bay during the fall cruise. A general analytical precision of 5 percent (relative std. dev.) was observed for the components methane, ethane, and ethene, whereas the remaining

components were found in concentrations too low to provide meaningful estimates of precision. Subsequent measurements made in July 1976 show that precision for propane and propene was also near 5 percent; again butanes were present at or below their detection limit.

PHYSICAL SETTING AND HYDROGRAPHY

Bristol Bay is a broad shelf region located in the southeastern Bering Sea. It is bounded on the south by the Alaska Peninsula, on the west by the shelf break, and on the north by the Alaska Coast (Schumacher et al. 1979). Freshwater influx, originating primarily from the Kuskokwim and Kvichak watersheds, averages 47 km³ annually (Kinder 1977). Ice covers approximately 60 percent of Bristol Bay between the months of December through April (Schumacher et al. 1979).

There are three principal water masses that have been identified in Bristol Bay (Coachman and Charnell 1977). The warmest and most saline water is found along the outer southern shelf (cf. Fig. 2, Coachman and Charnell 1977). Between this water mass and the 50-m isobath is the middle shelf water, which is usually stratified thermally in summer. Bottom temperatures of -1 C are not uncommon. The coastal water ($z < 50$ m) is characterized by the lowest salinity and may or may not be stratified, depending on the season, depth of water, and freshwater influx.

Currents over the shelf are generally weak (Kinder and Coachman 1978, Coachman and Charnell 1979). Along the outer shelf, in summer, current trajectories generally trend northwest at speeds of approximately 5 cm/sec. Across the inner shelf, currents are weak (~ 1 cm/sec) and variable.

Frontal structures have been observed along the 50-m and 100-m isobaths, which hydrographically separate the various water domains. Development of the front in shallow water appears to be related to the onset of stratification and tidal mixing (Schumacher et al. 1979).

RESULTS

Bristol Bay was sampled for dissolved LMW hydrocarbons in September and October 1975 and in June and July 1976. Vertical profiles were made at each of the stations shown in Fig. 27-1, although instrumental difficulties and weather forced curtailment of sampling at a few stations. To document temporal changes occurring over tidal frequencies, time-series

Figure 27-1. Location of stations occupied in Bristol Bay in Sept.-Oct. 1975 and July 1976. The solid lines show the vertical sections along which the distribution of properties is discussed. Depth contours are in fathoms.

measurements (24 hours and 36 hours) were made at Stations Ebb 37 and 46 in September and October, 1975. In addition to the sampling for dissolved hydrocarbons, measurements of salinity, temperature, and concentrations of suspended matter were also made at each of the stations shown (Feely and Cline 1977).

Methane

Methane is the dominant dissolved LMW hydrocarbon and its distribution reflects both seasonal and spatial source patterns. The distributions of dissolved methane in surface and near-bottom waters are shown in Figs. 27-2a and b for September and October 1975. Except for a localized source near Port Moller,

surface concentrations of methane were near values expected from the saturation of air. Assuming a methane partial pressure of 1.4 ppm(v) (Ehhalt 1974) and a mean surface temperature and salinity of 7°C and 31‰ (Kinder and Schumacher, Chapter 4, this volume), the equilibrium solubility concentration of CH_4 is 53 ± 3 nl/l (STP) (Yamamoto et al. 1976). The plume of methane observed south of Cape Newenham may have arisen from the Nushagak and Kvichak Rivers, although the small enrichment noted ($75 \text{ nl/l} < \text{CH}_4 < 94 \text{ nl/l}$) is near the ambient noise level when variability in time and space is taken into account. These data, however, do not rule out a contribution from bottom sediments in the Kuskokwim Delta.

Figure 27-2. Surface (a) and near-bottom (b) distributions of dissolved methane (nl/l, STP) in Sept.-Oct. 1975. Near-bottom samples were taken within 5 m of the bottom.

The elevated concentrations near Port Moller presumably arise from the tidal discharge of methane-rich waters from the lagoon. It is interesting, however, that methane is contained inside the 50-m isobath. The lack of a definite plume trajectory either east or west also suggests that there was little or no persistent coastal current at the time of these measurements. This is in general agreement with circulation studies carried out in Bristol Bay over the past several years (Kinder and Schumacher, Chapter 5, this volume), in which wind forcing or other meteorological events may give rise to a coastal current. The strong lateral methane gradient in the absence of any significant mean flow also suggests that the previously described frontal system along the 50-m iso-

bath (Schumacher et al. 1979) inhibits a horizontal flux of methane to the north.

The concentration of methane within 5 m of the bottom is shown in Fig. 27-2b. The most striking feature is the concentrated source over the outer shelf, referred to here as St. George Basin. Concentrations of methane near the bottom were in excess of 600 nl/l (STP) and represent approximately 12-fold supersaturation with respect to the atmosphere. The source of the methane is undoubtedly microbial, presumably arising from the activities of microorganisms at the sediment-water interface. According to Sharma (1979), bottom sediments along the outer shelf are fine grained, containing more than 0.5 percent organic carbon by weight. While these concentrations of carbon are not inordinately high, they are enriched compared to the inner shelf region, where concentrations range from 0.05 percent to 0.2 percent (Sharma 1979). The exception to this generalization is Kuskokwim Bay, where organic carbon concentrations range from 0.1 percent to 0.5 percent. These fine-grained sediments also are a likely source of methane, as mentioned earlier.

The methane distribution inside the 50-m isobath was vertically homogeneous. This distribution is expected if the bottom production rates are small compared to the rate of vertical mixing. Salt and heat were also vertically homogeneous in the coastal water, suggesting strong vertical mixing at the time of the measurements (Kinder and Schumacher, Chapter 4, this volume). For the sake of brevity, the distributions of salinity and temperature for that period will not be shown as they were similar to those observed by Schumacher et al. (1979) for the following year.

In Bristol Bay, the distribution of methane is controlled by several point sources and circulation. This is apparent in Fig. 27-3, where the vertical distribution of methane is shown along a north-south section between Nunivak Island and Unimak Pass (see Fig. 27-1; Sec. I). Methane in St. George Basin clearly originates from the bottom. While a weak mean current over the outer shelf (Kinder and Schumacher, Chapter 5, this volume) may transport methane to the northwest, this figure clearly shows the influence of vertical mixing on the distribution of methane. Because of increased vertical turbulence, methane-rich water was observed at the surface near Unimak Pass (Sta. 46). In contrast, the inner shelf region was nearly homogeneous with respect to dissolved methane, with concentrations near the equilibrium values.

In order to evaluate the short-term variability of methane, Stations 46 and Ebb 37 were sampled every 4 hours for 24 hours and 36 hours respectively. The results of the time-series measurements are

Figure 27-3. Dissolved methane (nl/l, STP) along a north-south section through Unimak Pass. Observations were made in Sept.-Oct. 1975.

shown in Figs. 27-4a and b. At the beginning of the observations (Sta. 46), the distribution of methane was vertically homogeneous, or nearly so. At approximately 4 hours, methane concentration in the surface layers increased to over 200 nl/l and remained high for the following 20 hours. Similar trends were observed at depth although there appeared to be a significant lag period (four to five hours). Because the only reasonable source of the methane is St. George Basin, it is assumed that during the observational period water was advecting south through Unimak Pass.

To the north at Station Ebb 37, just east of the Pribilof Islands, the concentration of methane at the surface and at depth was uniform over the entire observational period. This suggests that methane-rich water from St. George Basin did not move onto the shelf during this period. On the contrary, the mean current trajectory was probably parallel to the isobaths and would result in advection of dissolved methane toward the northwest (Coachman and Charnell 1979).

In contrast to conditions observed the previous fall, concentrations of all LMW hydrocarbons were elevated in the surface waters during July. Concentration of dissolved methane in the surface waters is shown in Fig. 27-5a. Port Moller once again was a significant source, as it was the previous fall. A concentration of methane greater than 1300 nl/l was observed at Station 28, or a value approximately 26 times the saturation amount. The source of this methane is believed to be sediments rich in organic matter inside Port Moller, where decomposition of vegetable matter could result in the vigorous produc-

Figure 27-4. Short-term variability of dissolved methane at stations 46 (a) and Ebb 37 (b) in Sept.-Oct. 1975.

tion of methane. In the absence of significant microbial oxidative processes, methane dissolved in the shallow brackish waters of the lagoon would be transported through the entrance by tidal pumping (Barsdate et al. 1974).

Because the air-sea exchange of methane is relatively slow, advection of methane-rich water could be traced east along the Alaska Peninsula for nearly 200 km. Although a cyclonic mean flow in the coastal water does not appear to dominate the shelf salt budget (Coachman and Charnell 1979), a weak coastal current exists along the peninsula (Kinder and Schumacher, Chapter 5, this volume).

The surface distribution of methane in July, not unlike temperature, delineates quite sharply the hydrographic domains over the shelf. Near the shore in water depths less than about 50 m, the concentrations of methane were everywhere greater than 100 nl/l and reflected increased production relative to the previous fall. The source of the methane is not known precisely, but would include coastal sources as well as *in-situ* production from both bottom sediments and the water column.

The Kuskokwim River may also be a significant source of methane as suggested by the high concentrations (> 400 nl/l) found near Cape Newenham. The Kuskokwim River was not sampled for dissolved hydrocarbons, but the Yukon River plume was sampled in July 1979 and found to contain concentrations of methane in excess of 2,000 nl/l. We would expect the lower reaches of the two rivers to be similar in dissolved methane content, since their lower drainage basins are similar. As mentioned

Figure 27-5. Surface (a) and near-bottom (b) distribution of dissolved methane (nl/l, STP) in July 1976. Near-bottom samples were taken within 5 m of the bottom.

above, however, the benthic production in Kuskokwim Bay must be considered on the basis of the available data.

Methane concentrations in the surface waters of the middle shelf during July ranged from 50-100 nl/l (Fig. 27-5a). Vertical stratification and low production of methane result in the observed distribution. The near-bottom distribution of methane is reflected in Fig. 27-5b. Previously identified sources near Port Moller and Cape Newenham are evident, although concentrations at depth are somewhat reduced. This observation would be expected if the rivers in the area are a significant source of the methane.

The bulk of the middle shelf water contained rather uniform levels of methane (50-100 nl/l), except for the outer shelf region north of Unimak Pass,

where concentrations in excess of 400 nl/l were found. These concentrations are slightly less than the values observed the previous fall (see Fig. 27-2b) and suggest that methane production is seasonal. Water temperature, carbon production rate, and quality of the organic matter would all be expected to strongly influence the production rate of methane, hence the *in-situ* concentration, if dispersive factors remain constant.

The influence of circulation and mixing on the distribution of methane is depicted in Fig. 27-6a, a zonal section through Bristol Bay (see Fig. 27-1; Sec. II). Methane is vertically homogeneous in the coastal water (Sta. 25A-17), but shows vertical structure over the middle shelf (Sta. 17-42). Surface waters in the middle shelf region are close to saturation with respect to methane in the atmosphere, whereas

Figure 27-6. (a) Vertical distribution of dissolved methane (n/l, STP) along Sec. II in Bristol Bay (see Fig. 27-1). (b) Vertical distribution of dissolved methane (nl/l, STP) along Sec. I terminating in Unimak Pass. Observations were made in July 1976.

concentrations increase to near 200 nl/l at depth. Analogous to the situation predicted for salt (Coachman and Charnell 1979), some of the methane found over the middle shelf probably originates by lateral diffusion from the outer shelf. Undoubtedly, there is an indigenous source, but its significance cannot be evaluated at this time.

A core of methane-rich water is evident at Station Ebb 48 (Fig. 27-6a). It presumably originates from the southeast in water depths near 120 m (see Fig. 27-6b; near Sta. 44). The core properties of the methane suggest a mean flow at depth toward the northwest. Current-meter observations of Kinder and Schumacher (Chapter 5, this volume) taken over the outer shelf for the period June-July 1976 show a net current speed of 1-2 cm/sec at 100 m to the northwest (see their Sta. BC-13B), supporting the observed methane trajectory.

The distribution of methane along a north-south section between Nunivak Island and Unimak Pass is shown in Fig. 27-6b. General structural features shown in Fig. 27-6a are preserved in this section, except for the complexity of multiple sources near Unimak Pass (Sta. 45-48). In contrast to the conditions observed in the previous year (see Fig. 27-3), there was a large accumulation of methane at depth south of Unimak Pass. The highest concentration of methane (630 nl/l) was observed at 400 m (Sta. 48), rather deep penetration for methane presumably originating from shallow shelf waters. However, since the concentration of methane at 100 m was 550 nl/l, methane-enriched water found offshore probably originated from the broad shelf along the southern Aleutian Peninsula. Whereas the data in Fig. 27-6b suggested a northward movement of water through Unimak Pass during the observational period, it probably was not sustained for any significant period of time. This is exemplified by our observations of methane in the vicinity of Unimak Pass during September and October 1975 (see Fig. 27-4a).

LMW hydrocarbons

The concentrations of the C_{2+} hydrocarbons (C_2 to C_4) are governed largely by seasonal processes involving biological activity or increased levels of insolation. Frontal dynamics appear to play a lesser role in controlling the distribution of these than for methane. A summary of the hydrocarbon concentrations (means and standard deviations) for the various hydrographic domains is shown in Table 27-1. For the purpose of clarity and to delineate possible source regions, the data were organized according to hydrographic domains as described by Coachman and Charnell (1977). Distinctions are also made between

surface waters and near-bottom waters.

Because of analytical problems encountered during the fall cruise, the concentrations of ethene and propene include those of ethane and propane. Many observations conducted in the coastal waters of Alaska show that the concentration of ethene is approximately three times that of ethane (see Fig. 27-8). The relationship of propene and propane is similar.

Since concentrations of butanes (iso- and *n*-) were always below the detection threshold of 0.05 nl/l, they are not included in Table 27-1. They do, however, occur in measurable amounts near well-defined gas seeps in Norton Sound and Cook Inlet (Cline and Holmes 1977, Cline 1977).

Both ethene and propene show a significant seasonal signature in surface waters. For example, the surface distribution of ethene during the summer of 1976 is shown in Fig. 27-7. Localized sources of ethene are not well defined, although the eastern portion of Bristol Bay appeared to be more productive than the outer shelf region in July. The distribution of ethene, like methane, does not correlate well with the hydrographic parameters. The reason for this probably lies in the relative rates of production and the decoupling of ethene production from hydrographic domains: water-column production of ethene is not influenced strongly by the 50-m frontal system, but is largely controlled by photochemical and biochemical processes in the surface and near-bottom waters.

During the fall, in all hydrographic domains, the concentration of ethene was less than 1 nl/l, increasing to concentrations greater than 2 nl/l in summer. A similar trend was noted for propene, although the concentrations were systematically less.

As expected, the vertical distribution of the C_1 - C_4 hydrocarbons in the coastal domain is invariant, although the seasonal component is evident in the depth-averaged mean concentrations (Table 27-1). This fact suggests that hydrocarbon production occurs in the water column or possibly at the sediment-water interface. If hydrocarbon production occurs principally at the sediment-water interface, tidally induced turbulence appears to be strong enough to homogenize the coastal water mass. A similar phenomenon obtains for heat and salt (Kinder and Schumacher, Chapter 4, this volume).

In the middle shelf domain ($50\text{ m} < z < 100\text{ m}$), the influence of both surface production and vertical stability come into play. Methane showed little seasonal difference (average), suggesting that water-column production was minimal and whatever vertical structure was present can be attributed to changes

TABLE 27-1

Average surface (a) and near-bottom (b) concentrations (nl/l, STP) of methane, ethane, propane, and propene for various water depth intervals.

CRUISE	DOMAIN	METHANE		ETHANE		ETHENE ¹		PROPANE		PROPENE ¹	
		Mean	Range	Mean	Range	Mean	Range	Mean	Range	Mean	Range
Sept.-Oct., 1975	Coastal (<50 m)	a	64	45-94	--	--	0.9	0.3-1.7	--	0.5	0.2-1.1
		b	59	45-98	--	--	1.0	0.7-1.8	--	0.4	0.1-0.6
	Middle Shelf (50-100 m)	a	60	42-83	--	--	0.8	0.3-1.6	--	0.4	0.1-1.4
		b	99	65-163	--	--	1.7	1.2-2.7	--	0.6	0.3-1.3
	Outer Shelf (100-200 m)	a	76	40-200	--	--	0.5	0.2-0.8	--	0.3	0.3-0.4
		b	380	100-615	--	--	1.1	0.7-1.6	--	0.3	0.2-0.4
	Coastal (<50 m)	a	112	74-153	0.9	0.6-1.5	3.8	3.0-4.7	0.4	1.4	1.0-2.5
		b	114	73-153	1.0	0.5-2.5	3.4	2.3-4.4	0.4	1.2	0.7-1.6
	Middle Shelf (50-100 m)	a	85	52-134	0.6	0.3-1.5	2.9	1.9-4.7	0.3	1.1	0.6-1.7
		b	115	62-165	1.3	0.5-2.5	2.2	1.1-4.0	0.5	0.5	0.2-1.0
June-July, 1976	Outer Shelf (100-200 m)	a	140	53-276	1.1	0.4-2.1	2.3	1.8-2.8	0.4	0.7	0.5-1.1
		b	269	164-440	0.9	0.6-1.1	1.2	0.8-1.8	0.3	0.3	0.1-0.9

¹ Due to analytical difficulties encountered during the Sept.-Oct. 1975 cruise, concentrations of ethene and propene include ethane and propane respectively.

in vertical stability. Because of vertical stratification, concentrations of methane below the thermocline are somewhat higher than in the surface layers, although not significantly so in the summer of 1976 (Table 27-1). This observation suggests a small benthic source.

In fall the highest concentration of ethene was found at depth, whereas in summer the concentration was higher in the surface layers. This difference is attributed to increased biological or induced photochemical production in the surface layers. Similar trends are apparent in the distribution of propene, but because of lower concentrations (i.e., lower production rate), seasonal and spatial variations are more obscure and probably not statistically significant.

The salient hydrocarbon features characteristic of

the outer shelf domain ($100\text{ m} < z < 200\text{ m}$) were the high concentration of methane in the near-bottom waters and a relative increase in the concentration of ethane and propane compared to that of ethene and propene. The ethane/ethene ratio was significantly larger in the bottom waters of the outer domain than observed elsewhere in Bristol Bay. This relationship is attributed to the influence of organic-rich, possibly anoxic sediments near the sediment-water interface that modify the normal alkane/alkene ratio. For example, the dominant C_{2+} hydrocarbons in the anoxic waters of the Black Sea are the alkanes, indicating that the production of biological hydrocarbons under anoxic conditions tends toward the more reduced alkanes (Hunt 1974).

Figure 27-7. Surface distribution of dissolved ethene (nl/l, STP) in July 1976.

DISCUSSION

Source-Composition Relationships

Low molecular weight aliphatic hydrocarbons are ubiquitous in the surface waters of the world oceans (Swinerton and Lamontagne 1974). Methane, the most abundant hydrocarbon, is generally at or near saturation in offshore surface waters, increasing in concentration in coastal and estuarine waters (Swinerton and Lamontagne 1974, Brooks and Sackett 1973, Scranton and Farrington 1977). Supersaturation of methane in coastal waters is largely attributed to methanogenesis occurring in bottom sediments (Oremland 1975), or similar processes occurring within suitable microenvironments (e.g., zooplankton, fecal matter, etc.) in the water column (Scranton and Farrington 1977). In localized areas of the Gulf Coast, gas seeps, underwater gas venting, and production activities associated with gas and oil development have led to significant enrichments of both methane and its natural saturated homologs in the water column (Brooks and Sackett 1973, Bernard et al. 1976, Brooks and Sackett 1977).

The source of C_{2+} hydrocarbons in pristine marine waters is not well understood. Our work in the coastal waters of Alaska clearly demonstrates that ethane, ethene, propane, and propene are limited to surface and shelf waters. Vertical profiles taken in the Alaskan Trench showed a rapid decrease in the concentrations of these hydrocarbons below 200 m. These observations and those of others (Swinerton and Lamontagne 1974, Brooks and Sackett 1977) seem to indicate that the higher homologs of methane, including the alkenes, are derived from biological activity or perhaps photochemical reactions involving organic precursors in surface waters (R. Zika, Univ. of Miami, personal communication). These conclusions are supported by the laboratory studies of Wilson et al. (1970), and the *in-situ* observations of Swinerton et al. (1977). In the latter report, correlations were observed between hydrocarbon production, light intensity, dissolved organic carbon, and chlorophyll *a*.

To effectively distinguish petroleum-derived hydrocarbons from biologically produced components, compositional or isotopic (^{14}C , ^{13}C) patterns must be characterized. One of the objectives of this report is to suggest several characteristic compositional patterns that might be useful in identifying the source of hydrocarbons. One such parameter is the ratio of methane to ethane plus propane (Brooks and Sackett 1973). The other, which will be developed here, is the ethane/ethene ratio. These two hydrocarbon species are particularly useful in a diagnostic sense because ethene is abundant in biological systems and

absent in thermogenic sources. Conversely, ethane is relatively abundant in natural gas and petroleum.

In Fig. 27-8 the relationship between ethane (C_2) and ethene ($C_{2:1}$) is shown in two distinct hydrographic domains of Bristol Bay during July 1976. The relationship between ethane and ethene in surface waters (open circles) is shown by line (b). Linear regression of the observations yields the equation:

$$[C_2] = 0.28[C_{2:1}] - 0.20 \quad (r = 0.79).$$

Analyses made within 5 m of the bottom (solid circles) in the middle shelf domain also plot along this line, suggesting a common mechanism. In July, the water column was vertically stratified and rather turbid (Feely and Cline 1977); thus it is expected that photochemical reactions in the surface layers would have little effect on the distribution of hydrocarbons at depth.

As the deeper waters of St. George Basin are approached, the concentration of ethane increases relative to that of ethene (Fig. 27-8; line (a)). The equation of this line is:

$$[C_2] = 0.36[C_{2:1}] + 0.54 \quad (r = 0.88),$$

which reflects a higher production rate of ethane than of ethene. For comparative purposes, surface concentrations of ethane and ethene from Unimak Pass and waters near Izembek Lagoon also plot along this line.

Surface and middle shelf domain waters show a characteristic relationship in the relative abundances of ethane and ethene. In general, the concentration of ethene exceeds that of ethane by a factor of three

Figure 27-8. Relationship between the concentration of ethane and ethene for surface waters (open circles) and near-bottom waters (solid circles). Surface and near-bottom waters in the middle shelf region plot along line (b), whereas bottom waters from St. George Basin fall along curve (a). Surface waters from Unimak Pass and Izembek Lagoon also fall along curve (a). Transitional waters between St. George Basin and the middle shelf are outlined by the dashed line.

in well-oxygenated surface waters, like most Alaskan shelf areas studied (Cline et al. 1978). In the deeper water of St. George Basin ($z > 100$ m), the increase in the concentration of ethane is presumably related to the increased carbon flux to the sediments (Sharma 1979) and a concomitant drop in redox potential. Decaying plankton lying on the bottom (R. Reeburgh, Univ. of Alaska, personal communication) would be expected to give rise to an anoxic stratum at the bottom. This condition would favor increased production of ethane as was shown to occur in the anoxic waters of the Black Sea (Hunt 1974). It is not proposed that the near-bottom waters of St. George Basin are anoxic, only that microenvironments or microstrata near the bottom represent a significant source of ethane to the water column. Because the sediments of Izembek Lagoon are rich in organic matter (Barsdate et al. 1974), and are presumably anoxic, it is not surprising to find elevated concentrations of ethane near the entrance of the lagoon.

The relatively high concentrations of ethane found in the surface waters of Unimak Pass are the result of strong vertical mixing from below, but the possible influence of shipping, fishing, and transportation activities on the hydrocarbon distribution cannot be assessed at the present time.

As stated previously, hydrocarbons of different sources reflect distinct compositional patterns. Gaseous hydrocarbons generated in well-oxygenated waters or from shallow horizons in bottom sediments are usually characterized by high concentrations of methane (Brooks and Sackett 1973) and relatively high concentrations of alkenes (Bernard et al. 1978). Hydrocarbons derived from thermal processes are characteristically enriched in C_{2+} alkanes relative to methane and devoid of the alkenes (Frank et al. 1970, Clark and Brown 1977). Frank et al. (1970) have used these compositional characteristics to propose the ratio $[C_1]/[C_2] + [C_3]$ as a possible indicator of a source. Ratios in excess of 500 suggest a biological source, while ratios less than 50 indicate a thermogenic origin.

In an attempt to elucidate the origin of gaseous hydrocarbons, the ethane/ethene $[C_2]/[C_{2:1}]$ ratio was plotted against the methane/ethane plus propane $[C_1]/[C_2] + [C_3]$ ratio. Fig. 27-9 shows such a plot for distributions in Bristol Bay and includes two contrasting marine environments. Comparisons are made with the hydrocarbon distributions in the vicinity of the Norton Sound gas seep (Cline and Holmes 1977) and with the anoxic waters of the Black Sea (Hunt 1974). Including all the analyses from Bristol Bay, the $[C_1]/[C_2] + [C_3]$ ratio ranges from 30 to 500, while the $[C_2]/[C_{2:1}]$ ratio ranges between 0.1 and

Figure 27-9. Compositional hydrocarbon trajectory diagram for Bristol Bay. For comparison, the compositional fields of the Norton Sound gas seep (Cline and Holmes 1977) and the anoxic waters of the Black Sea (Hunt 1974) are indicated. Mixing of Bristol Bay hydrocarbons with various thermogenically derived hydrocarbon mixtures results in a family of mixing trajectories, two of which are shown here (see text). Along one of the dry gas trajectories, the fraction (%) of ambient water is given, assuming that the hydrocarbons in the source are each about 100 times the ambient levels.

1. These compositional ratios obtain for most of the Alaskan shelf waters and are probably typical of high latitude, pristine coastal waters. On the other hand, anoxic waters in the Black Sea (Hunt 1974) reflect a narrow compositional field characterized by high concentrations of methane, $[C_1]/[C_2] + [C_3] \cong 500$, and relatively high concentrations of ethane, $20 < [C_2]/[C_{2:1}] < 50$. Provisionally, it is assumed that the Black Sea represents a model of the hydrocarbon composition to be expected from anoxic marine sources.

At this point it becomes useful to consider the compositional trajectories that might be observed if gases of thermogenic origin were mixed with biologically derived hydrocarbons. To accomplish this, four distinct petroleum/natural gas sources were considered from an analysis of gas and oil well data (Moore et al. 1966). The equations that govern the mixing trajectories in the $[C_1]/[C_2] + [C_3]$ and $[C_2]/[C_{2:1}]$ plane are:

$$\frac{[C_1]}{[C_2] + [C_3]} = \frac{[C_1]_A + ((1-x)/x)[C_1]_T}{[C_2]_A + [C_3]_A + ((1-x)/x)[C_2]_T + [C_3]_T} \quad (1)$$

$$\frac{[C_2]}{[C_{2:1}]} = \frac{[C_2]_A}{[C_{2:1}]_A} + ((1-x)/x) \frac{[C_2]_T}{[C_{2:1}]_A} \quad (2)$$

The fraction of ambient water is x , whereas concentrations of the individual hydrocarbons are shown in brackets. The two sources, between which mixing

is assumed to occur, are indicated by the subscripts A (ambient) and T (thermogenic source). The implicit assumptions are that the source ratios are constant and that the concentration of ethene in the natural gas source is zero.

To develop possible mixing scenarios, it was necessary to evaluate the possible range of LMW hydrocarbon mixtures that might occur as the result of offshore production. The $[C_1]/[C_2]+[C_3]$ frequency diagram for 366 terrestrial gas wells (Moore et al. 1966) was plotted, and by this means three discrete compositions, identified on the basis of their $[C_1]/[C_2]+[C_3]$ ratio, were defined, ranging from what is described as a "very" dry gas (methane rich) to a "typical" wet gas (methane poor). Some of the wet gases were associated with petroleum. A summary of the calculations is shown in Table 27-2.

TABLE 27-2.

Mean mole fraction of methane, ethane, and propane, calculated for three arbitrarily defined natural gas fractions (Moore et al. 1966). Number of samples used in each statistic calculation is shown in column 5. Standard deviation about the mean is shown in parentheses.

Mole Fraction			$[C_1]$ $[C_2]+[C_3]$ ⁿ		Definition
C ₁	C ₂	C ₃			
.79(.11)	.075(.04)	.033(.026)	7	313	Wet Gas
.96(.02)	.011(.005)	.002(.002)	61	47	Dry Gas
.98(.02)	.001(.0006)	Tr	672	6	"Very" Dry Gas

The actual concentration ratio observed in the water will depend on the component partial pressure (or mole fraction), the Bunsen coefficient, which depends upon salinity and temperature, and the depth of water at which gas injection occurs. Implicit in the previous statement is that equilibrium is achieved between the gaseous and aqueous phases and that the rate of solution of the gases is a function of the thickness of the stagnant film boundary layer (Broecker and Peng 1974). Actually, equilibrium is probably not achieved. However, because the Bunsen coefficients and diffusion coefficients are similar for methane, ethane, and propane, the solubility ratio will not be significantly different from the equilibrium ratio. If these minimum conditions hold, the following expression relates the component partial pressure in the gas phase to the equilibrium solubility ratio:

$$\frac{[C_1]'}{[C_2]'+[C_3]'} = \frac{\beta_1 D_1 p_1}{\beta_2 D_2 p_2 + \beta_3 D_3 p_3} \quad (3)$$

where β_i , D_i , and p_i represent the Bunsen coefficient, diffusion coefficient, and mole fraction of (1) methane, (2) ethane, and (3) propane. Assuming a mean temperature and salinity of 5 C and 30‰,o,

$$\begin{aligned} D_1 &= 0.85 \times 10^5 \text{ cm}^2/\text{sec} & \beta_1 &= 0.04 \text{ ml CH}_4 \text{ (STP)/ml H}_2\text{O} \\ D_2 &= 0.69 \times 10^5 \text{ cm}^2/\text{sec} & \beta_2 &= 0.06 \text{ ml C}_2\text{H}_6 \text{ (STP)/ml H}_2\text{O} \\ D_3 &= 0.55 \times 10^5 \text{ cm}^2/\text{sec} & \beta_3 &= 0.06 \text{ ml C}_3\text{H}_8 \text{ (STP)/ml H}_2\text{O} \end{aligned}$$

Bunsen coefficients (β_i), corrected for the "salting out effect" and the molecular diffusion coefficients (D_i), not corrected for the ionic strength of seawater, were estimated from the data of Bonoli and Witherspoon (1968). The solubility ratio, $[C_1]'/[C_2]'+[C_3]'$, for gas well data calculated from equation 3, is shown in Table 27-2 and is also shown as mixing end members in Fig. 27-9. Because it has been assumed that the concentration of ethene is zero (i.e., $[C_2]'/[C_{2:1}]' \rightarrow \infty$), the mixing end members are located at the extreme right margin.

The remaining end member we wish to consider is a wet gas associated with petroleum. For this purpose, the volatile fraction from the reservoir fluid of the Sadlerochit formation, Prudhoe Bay, was selected (Anon. 1971). The so-called volatile fraction, C₁ to C₄ hydrocarbons plus air gases, was normalized to 100 percent and the partial pressures of methane, ethane, and propane were calculated. As in the example given earlier, a gas of this composition is equilibrated with a parcel of Bristol Bay water. Table 27-3 gives the mole fraction composition of the gas phase and the resulting equilibrium solubility ratio. Bunsen coefficients and diffusion coefficients were the same as before.

TABLE 27-3.

Mole fractions of methane, ethane, and propane from the reservoir fluid of the Sadlerochit formation, Prudhoe Bay (Anon. 1971), and the resulting equilibrium solubility ratio. Gas phase was assumed to contain only C₁-C₄ hydrocarbons, nitrogen, carbon dioxide, and helium.

Mole Fraction			$[C_1]'/[C_2]'+[C_3]'$
C ₁	C ₂	C ₃	
0.44	0.051	0.030	3.7

As expected, the gas is relatively rich in C₂₊ hydrocarbons and its composition is similar to that calculated earlier for a wet gas derived from gas wells. Of the gases present in the normalized mixture, methane, ethane, and propane constituted 81.6 percent by vol-

ume, C_4 hydrocarbons amounted to 3.4 percent, CO_2 was 14.2 percent, and the remaining 0.7 percent was divided between N_2 and He. Having estimated some possible petroleum and natural gas end members, trajectories that would result from mixing these end members with Bristol Bay water can be examined. The mixing trajectory resulting from the injection of a typical dry gas will be considered first.

To simplify the calculation, it is assumed that ambient Bering Sea water contains approximately the following concentrations of LMW alkanes: $[C_1] = 100$ nl/l, $[C_2] = 1$ nl/l, $[C_{2:1}] = 3$ nl/l, and $[C_3] = 0.4$ nl/l, giving a $[C_1]/[C_2]+[C_3]$ ratio of 71 and a $[C_2]/[C_{2:1}]$ ratio of approximately 0.33. It is apparent that this locus may be biased toward low values of the $[C_1]/[C_2]+[C_3]$ ratio (Fig. 27-9). If we assume that the local source of hydrocarbons is 100 times the ambient levels, the resulting mixing trajectory, computed from equations 1 and 2, is shown in Fig. 27-9. The values above the solid circles represent the percentage of ambient water at those points. Mixing a gas of this composition with the ambient water results in little change in the $[C_1]/[C_2]+[C_3]$ ratio, but a large shift in the $[C_2]/[C_{2:1}]$ ratio. This results directly from the assumption that the concentration of ethene was zero in the natural gas source. It is interesting that the mixing line passes through the compositional field observed in the region of the Norton Sound gas seep (Cline and Holmes 1977). Conclusions as to the source of thermogenic hydrocarbon gases in Norton Sound on the basis of this diagram would be misleading, however, since the ambient $[C_1]/[C_2]+[C_3]$ ratio in Norton Sound is significantly higher (~ 500) than the average value assigned for Bristol Bay (~ 71).

One additional feature of the compositional field diagram (Fig. 27-9) is that the mixing line between two defined sources is invariant with respect to the source concentrations of the hydrocarbons, as long as the ratios are fixed. This means that regardless of the source strength, compositional changes due to mixing will occur along the mixing line between the two sources. With the relative concentrations (i.e., 100:1) chosen in the above example, seepage or leakage of a dry gas in Bristol Bay would be observable to approximately 99 percent dilution. If the relative concentration ratio is increased to 1000:1, the effect is observable to 99.9 percent dilution.

Finally, we calculate the hypothetical mixing line, assuming a wet gas composition similar to that found in the Sadlerochit formation. This trajectory is also reflected in Fig. 27-9 and lies significantly below the previously calculated dry gas curve.

General conclusions drawn from Fig. 27-9 are that

typical wet and dry gases should be readily distinguishable from biologically produced hydrocarbons using a combination of the $[C_1]/[C_2]+[C_3]$ and $[C_2]/[C_{2:1}]$ ratios. The composition of LMW hydrocarbon mixtures characterized by $[C_1]/[C_2]+[C_3] < 20$ and $[C_2]/[C_{2:1}] > 100$ strongly suggests that they are thermally derived. The exception is a methane-rich dry gas, such as that produced currently from several wells in upper Cook Inlet (Kelly 1968). These gas wells contain methane in excess of 98 mole percent, only traces of ethane, and no measurable propane. An injected gas of this composition, in all likelihood, will not be distinguishable from the suite of hydrocarbons formed biologically under anoxic conditions (e.g., in sediments, lagoon environments, anoxic waters). Consequently, a dry gas of this composition would be interpreted as biogenic on the basis of its $[C_1]/[C_2]+[C_3]$ and $[C_2]/[C_{2:1}]$ ratios.

Additional useful tracers include the $[C_2]/[C_3]$ ratio (Nikonov 1972) and the $\delta^{13}C$ composition of the seep methane (Brooks et al. 1974, Bernard et al. 1976). In the latter study, Bernard and coworkers have plotted the $\delta^{13}C$ of the dissolved methane against the $[C_1]/[C_2]+[C_3]$ ratio for a number of gas seeps investigated along the Texas shelf. Biogenic methane was found to be isotopically light (-60‰ to -70‰ vs. PDB), whereas methane derived from thermogenic sources was significantly heavier at -40‰ to -50‰ . In a similar study, Kvenvolden et al. (1979) found the pore fluid methane at the locus of the seep in Norton Sound to be isotopically heavy at -36‰ , suggesting a thermal source.

There is no doubt that the isotopic composition of dissolved methane is an excellent diagnostic parameter in identifying the origin of natural gas. However, usually the concentration of methane is so low as to preclude the use of isotopic fingerprinting, except where gas venting is vigorous (e.g., Bernard et al. 1976). At low concentrations, the compositional ratios should be useful in identifying sources of hydrocarbons and their spatial trajectories. Because these hydrocarbons are dissolved, they readily identify the trajectories of other dissolved components, which may possess greater toxicity (e.g., benzene).

In the following discussion, attention will be given to the dispersion field of LMW hydrocarbons originating from a localized source and some estimates of the areal diffusion and advection scales applicable to Bristol Bay.

Dispersion model

The major advantages of using LMW hydrocarbons as *in-situ* tracers of petroleum are related to their relatively high abundance in crude oil and natural gas

and the low concentration levels at which they can be measured. In the event of a spill, well blowout, or subsurface seep, it becomes useful to define the trajectory and areal impact of the dissolved and suspended hydrocarbons. Because the LMW hydrocarbons are moderately soluble, these compounds become useful tracers of the soluble fraction. It is not the purpose of this report to model actual gas or oil seeps in Bristol Bay, since none were found, but rather to provide general guidelines on the usefulness of LMW hydrocarbons as tracers of the dissolved fractions of petroleum.

For simplicity we assume that the petroleum source is continuous and steady in time, and results in a vertically homogeneous distribution within a selected layer (e.g., surface mixed layer). The model describing these minimum conditions is a two-dimensional advection-diffusion equation given by Csanady (1973):

$$\frac{\partial}{\partial y} [K_y \frac{\partial C}{\partial y}] - U \frac{\partial C}{\partial x} - kC = 0, \quad (4)$$

where C is the concentration of the dissolved hydrocarbon, x and y are space coordinates, K_y is the scale-dependent horizontal eddy diffusivity, U is the mean horizontal velocity in the x -direction, and k is a first-order decay constant. In this case, we will use the first-order decay term to describe the air-sea exchange process and assume biological oxidation and production to be negligible. Because the source of hydrocarbons is presumably small, the horizontal eddy diffusivity depends on the mixing scale (Okubo 1971). The solution to equation 4 for a line source is given by Csanady (1973) and is reproduced here without derivation:

$$\bar{C}(x,y) = \frac{1}{2}\bar{C}_0 \exp(-kx/U) [\operatorname{erf}((L/2+y)/\sqrt{2\sigma_y}) + \operatorname{erf}((L/2-y)/\sqrt{2\sigma_y})] \quad (5)$$

where $\bar{C}_0$ is the vertically integrated concentration at $x = 0$, L is the length of the line source, and σ_y is related to K_y through the equation $K_y = (U/2)(d\sigma_y^2/dx)$. Equation 5 is given in terms of the concentration at $x = 0$ rather than the production rate. This was done because the input rate for a seep would be difficult to quantify. Moreover, to our knowledge the production rate of known submarine gas seeps has never been documented. Thus, the model is expressed in terms of the concentration field, which should be easily measured. To simplify the discussion in terms of maximum trajectory scales, only the cen-

terline distributions ($y = 0$) will be discussed. Hence:

$$\bar{C}(x,0) = \bar{C}_0 \exp(-kx/U) [\operatorname{erf}(L/\sqrt{8\sigma_y})] \quad (6)$$

Using empirical relationships presented by Okubo (1971), where $\sigma_{rc}^2 = 0.0108t^{2.34}$, $U = x/t$, and $\sigma_{rc}^2 = 2\sigma_x\sigma_y$, equation 6 becomes

$$\bar{C}(x,0) = \bar{C}_0 \exp(-kx/U) [\operatorname{erf}(L/0.208(x/U)^{1.17})] \quad (7)$$

after substitution.

In Bristol Bay, as in any body of water, the source of gas may be at the surface or at depth. To discuss dispersion scales under both these conditions we will assume that (1) the source is at the surface and is consequently influenced by air-sea exchange and (2) the source is at depth and physically isolated from the surface by a strong pycnocline ($k = 0$). We further assume that the component is biologically unreactive, which is a reasonable assumption at low concentrations. Methane will be selected as the model component but any of the LMW hydrocarbon species would suffice. Equation 5 is valid for all dissolved hydrocarbon species that possess chemical and biochemical reactivities similar to that of methane.

Under the assumptions of the model, methane injected at the surface will be dispersed by diffusion and advection, and lost from the system through air-sea transfer. The flux of methane across the air-sea boundary can be described by the stagnant film boundary layer model (Broecker and Peng 1974),

$$F_1 = (D/h)(\bar{C}_1 - C_1'), \quad (8)$$

where D is the molecular diffusion coefficient, h is the thickness of the stagnant film, $\bar{C}_1$ is the average concentration in the mixed layer, and C_1' is the equilibrium solubility concentration. The thickness of the molecular film is a function of sea-surface roughness or wind velocity (Emerson 1975). From the derivation of Fick's second law, it can be readily shown that transport across the sea surface is:

$$T_1 = (D/\Delta z \cdot h)(\bar{C}_1 - C_1'), \quad (9)$$

where Δz is the depth of the surface mixed layer. Thus from equation 9 and by analogy to equation 4:

$$k = D/\Delta z \cdot h. \quad (10)$$

In order to assess the importance of air-sea exchange on the methane dispersion scale, the value of k must be estimated. To accomplish this, nominal seasonal values for surface temperature and mean sca-

lar surface winds were used. All other parameters were computed from these data. Parameters used in the calculation of k are summarized in Table 27-4.

TABLE 27-4

Parameters used in the estimation of the first order decay constant (equation 10) governing the escape rate of methane across the sea surface.

	Temp. ^a °C	Wind Speed ^b m/sec	D_1 ^c cm ² /sec	Δz ^a m	h ^d μm	k /sec
Summer	10	6.3±0.8	1.1×10 ⁻⁵	20	80	6.6×10 ⁻⁷
Winter	0	9.9±0.5	0.65×10 ⁻⁵	75	15	5.8×10 ⁻⁷

^aKinder and Schumacher, Chapter 4, this volume

^bBrower et al., 1977

^cBonoli and Witherspoon, 1968

^dEmerson, 1975

Mean monthly wind velocities were averaged for the months of May-August (summer) and November-February (winter). Diffusion coefficients were taken from the data presented by Bonoli and Witherspoon (1968) and corrected to 0.5M NaCl solutions according to their equation. Thickness of the molecular film as a function of wind velocity was read from the graph presented by Emerson (1975). Because the effects of sea-surface roughness on the stagnant film thickness (h) and mixed layer depth on the rate constant (k) are seasonally compensating, a mean value of 6.2×10^7 /sec was adopted in the calculation (Table 27-4). For comparison, this value is approximately one-half the value calculated for the surface waters of Walvis Bay by Scranton and Farrington (1977).

The solution of equation 7 requires that an appropriate source dimension (L) and mean velocities (U) be chosen. Equation 7 is the solution for a line source of length L ; however for our purpose here, we will assume that the source is distributed along L rather than an extended point source. The results are equivalent at distances greater than $10 L$ (personal communication, J. Lavelle, PMEL). The length of the source was set arbitrarily at 1 km, which is probably appropriate for a surface spill, but would be excessive for small subsurface seeps.

Current trajectories and mean current velocities for Bristol Bay are known (Kinder and Schumacher, Chapter 5, this volume). Typical values for the middle shelf are 2 cm/sec, whereas velocities near the slope are in the neighborhood of 10 cm/sec. These

values were selected for simulation of the dispersion field, although velocities greater than 10 cm/sec may occur during storms (Kinder and Schumacher, Chapter 5, this volume). Of course, higher velocities are observed at tidal frequencies, but this effect is included as diffusive transport in equation 4 (Coachman and Charnell 1979).

The results of the simulation are reflected in Fig. 27-10 for the assumed velocity fields. Dashed lines show the centerline normalized concentration of methane in the absence of air-sea evasion ($k=0$). If it is assumed that the input rate along L increased the local concentration of methane to a factor of 100 (C/C_0) above background (i.e., 10,000 nl/l), methane could be observed down the axis of flow for a distance of 150 km in the surface layers and 400 km in the bottom waters, assuming a mean current flow of 10 cm/sec. Actual trajectory scale in the surface waters will probably be somewhat greater, because the air-sea transfer rate is proportional to the term $(\bar{C}_1 - C_1')$ (see equation 9). In effect, the transfer rate decreases as the concentration of methane in the

Figure 27-10. Relative decrease in the concentration of a dissolved hydrocarbon from a line source 1 km in diameter as estimated by a steady-state, horizontal diffusion/advection model. Surface ($k = 6.2 \times 10^7$ /sec) and near-bottom trajectories ($k = 0$) are considered for mean flows of 2 cm/sec and 10 cm/sec.

plume approaches the saturation value; in this case the equilibrium concentration would be approximately 50 nl/l. At reduced mean current velocities, lateral diffusion becomes more important, reducing the concentration of methane along the axis of the plume. At velocities of only 2 cm/sec, methane would be observable along the centerline of the plume for distances of 40 km and 80 km, depending on whether the injection was at the surface or at depth ($k = 0$). Clearly, small seeps can be traced for considerable distances, particularly if the injection occurs at depth, where hydrostatic pressure increases the local concentration.

At this point, it becomes pertinent to ask what hydrocarbon enrichment might be expected from a gas seep, natural or otherwise. A typical wet gas, as described earlier (Fig. 27-9), might be expected to contain mole fractions of methane, ethane, and propane of 0.80, 0.075, and 0.033 respectively. Assuming that the injection of a gas of this composition occurred at 50 m (ca. 5 atm hydrostatic pressure) and that the water column became saturated locally with respect to each of these gases, concentrations of methane, ethane, and propane would reach saturation values of 1.8×10^8 nl/l, 2.5×10^7 nl/l, and 1.0×10^7 nl/l, respectively. (It has been assumed that the effect of hydrostatic pressure on solubility is linear.) Assuming ambient levels of methane, ethane, and propane of 200 nl/l, 2 nl/l, and 1 nl/l (these values are about twice the actual mean values for the middle shelf region; see Table 27-1), the enrichment factors for methane, ethane, and propane become 1.0×10^6 , 1.2×10^7 , and 1.0×10^7 , respectively. Assuming that the injection of gas was at 50 m as above, these hydrocarbons would be observable along the axis of the plume for distances up to 1,000 km. Here a signal-to-noise ratio of approximately two is assumed. Actual longitudinal trajectories will depend on a number of factors: the scale of the seep, mean current velocities, diffusivity near the source, and biological oxidation processes in the water and at the surface of the sediments. Biological oxidation is likely to be much more important at the higher concentrations of hydrocarbons, thereby reducing the trajectory scale. Even in the surface layer, we estimate that LMW hydrocarbons could be traced for several hundreds of kilometers, depending on a number of factors. Among these are the state of the surface microlayer (wind velocity, surface organics, etc.), depth of the mixed layer (vertical dilution), microbial oxidation, and the mean current velocity.

While the above calculations are crude, they nevertheless serve to delineate the usefulness of LMW hydrocarbons as effective tracers over moderate space

scales. In the Norton Sound gas seep (Cline and Holmes 1977), the ethane plume was observable for at least 140 km downstream of the source for a source enrichment of about 20 above ambient. Referring to Fig. 27-10, and assuming a 2 cm/sec mean current velocity near the bottom (Muench et al., Chapter 6, this volume), the model predicts a longitudinal trajectory scale of 100 km ($k=0$). In this case, the model appears to underestimate the actual observed scale, possibly by as much as a factor of two. The reasons for the disparity are numerous; among them are variable gas seepage rate, fluctuating currents, inappropriate near-source diffusive scale, and poor characterization of the dimension of the ethane source. Given the hydrographic uncertainties and model assumptions, the prediction is probably as good as can be developed at this time.

SUMMARY

The concentrations of dissolved LMW hydrocarbons in Bristol Bay are relatively low when compared with other Alaskan shelf waters. Methane, the most abundant hydrocarbon, is near saturation concentration in surface waters, but increases significantly near lagoons along the Alaska Peninsula. There appears to be no major production of methane from the bottom sediments of the middle shelf but significant amounts are produced seasonally from the organic-rich sediments of St. George Basin. The distribution of methane is largely controlled by mean flow and frontal dynamics.

The C_{2+} fraction shows seasonal variability and appears to be regulated by biological processes, presumably microorganisms. As observed elsewhere in Alaskan coastal waters, the alkenes are more abundant than the alkanes of the same carbon number. The concentration of ethane, for example, has a linear relation to the concentration of ethene, which suggests a common precursor or source. Low concentrations of LMW hydrocarbons, a relatively high $[C_1]/[C_2]+[C_3]$ ratio of 30 to 500, and a $[C_2]/[C_{2:1}]$ ratio of less than 1.0, all suggest a biological source. Since no evidence was found in the LMW hydrocarbon fraction for the presence of petroleum-like hydrocarbons, we conclude that Bristol Bay is pristine.

Comparison of the hydrocarbon composition of typical natural gas and petroleum with those found in Bristol Bay shows that small amounts of thermogenically derived gases should be readily discernible in the water column, depending, of course, on the magnitude of the source. On the basis of the $[C_1]/[C_2]+[C_3]$ and $[C_2]/[C_{2:1}]$ ratios, a thermogenic

field applicable to Bristol Bay can be defined. In general, $[C_1]/[C_2]+[C_3]$ ratios less than 20 and $[C_2]/[C_{2:1}]$ ratios greater than 100 strongly suggest a thermogenic source. A very dry natural gas ($CH_4 > 98$ mole percent), similar to that currently being produced in Cook Inlet, represents an exception to the foregoing generalizations. In all likelihood, a methane-rich gas of thermogenic origin would appear to be of biological origin in the $[C_1]/[C_2]+[C_3]$ and $[C_2]/[C_{2:1}]$ compositional fields.

Finally, an estimate was made of the possible dispersion scale of LMW hydrocarbons arising from a line source, using a steady-state two-dimensional model. On the basis of typical mean velocities of 2 and 10 cm/sec with the inclusion of a sink term for air-sea exchange, it was estimated that dissolved LMW hydrocarbons would be identifiable in surface waters over distances of 40-150 km, assuming a local hydrocarbon concentration of 100:1 above ambient. At depth below a strong pycnocline, the horizontal mixing scale increases to 100-500 km. Actual trajectory scales expected in Bristol Bay will depend on a number of factors. Among them are the strength of the source, degree of solution, diffusive mixing scales, air-sea exchange, and biological oxidation rates.

In this report the usefulness of LMW hydrocarbons as a diagnostic indicator of petroleum hydrocarbon is described. Not only are these gases excellent tracers of dissolved or emulsified petroleum hydrocarbons, but their ease of measurement allows real-time observations to be made—a valuable adjunct to a monitoring program.

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Section V

Fisheries Oceanography

Felix Favorite, editor

Overview of Fisheries Oceanography

28

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ABSTRACT

In the past, specific fisheries oceanography investigations designed to explore or explain the complex interrelationships of fish and the various elements of the eastern Bering Sea environment have not been carried out in spite of the fact that early investigators consistently pointed out the inadequacies of casual environmental observations obtained during fishing cruises and recognized the need for more complete studies. Although today the acquisition and processing of environmental (physical and chemical) data are rapid and straightforward, using STD and autoanalyzer systems and shipboard computers, integrated studies are still rare. This overview puts the subsequent chapters in this section in the perspective of fisheries oceanography and also suggests that although annual stock assessments may satisfy management criteria, conservation practices will require information on the year-round distributions of juveniles as well as adults, multispecies interactions, and rather specific relations within the ecosystem. This is a large task, but more emphasis on fisheries oceanography studies will accelerate acquisition of the required knowledge.

INTRODUCTION

The rapacious exploitation of whales in the Bering Sea to obtain whale oil in the mid-19th century was accomplished without any consideration of conservation principles or of impact on the ecosystem; it is gratifying to see the extensive efforts funded by the Outer Continental Shelf Environmental Assessment Program (OCSEAP) to define conditions and processes in this area as in this century we seek to obtain petroleum from below the sea floor. Because of the purpose of the OCSEAP program, to assess real and possible effects of oil exploration and exploitation, many of the individual grants are largely disciplinary (e.g., physical oceanography, fishery biology, ichthyoplankton), and area specific (e.g., Norton Sound, Navarin Basin, St. George Basin, Bristol Bay Basin). The purpose of this section is to consider the continental shelf of the eastern Bering Sea as an integral unit and briefly evaluate environmental and biological continuities and relationships of certain fishes

throughout the area. Although we have far exceeded the five chapters called for in the initial guidelines for this study, the subject is still not adequately covered.

The term "fisheries oceanography" stems from fisheries hydrography, which originated at the turn of the century, and essentially embraced any and all conditions influencing fisheries. Since the word "ecology" apparently was avoided, one suspects that originally the term meant the study of those immediate conditions that would help to catch fish, or perhaps explain why fish were not caught, rather than the broad studies of the ocean or even very specific but basic research. The shift from fisheries hydrography to fisheries oceanography in mid-century was largely semantic, and did not represent a change in activities; those in the burgeoning field of ocean studies could not decide whether they would march under the banner of oceanography or oceanology. Such a discussion may seem trivial, but Chapman (personal communication) canvassed the world in an attempt to obtain an acceptable definition of fisheries oceanography, without success.

The actual scope of studies embraced by fisheries oceanography is virtually unlimited—from external forces, global or extraterrestrial, that alter the physics, chemistry, biology, or geology of the sea, to internal forces that redistribute, modify, consume, or guide this energy, including interactions of all life, life-support systems, and the effects of man. Currents, temperatures, and density have long been considered a part of fisheries studies. Although one may consider dismissing studies of nutrient chemistry because the absence or abundance of nutrients is reflected in primary (plant) production, without these studies there will be no evidence of why phytoplankton are scarce or abundant. In a study of pelagic fish one can argue that even primary production can be dismissed because it is reflected in sec-

ondary production (herbivores), but primary production cannot be ignored in discussing demersal or bottomfish, which consume benthos, whose development depends in large part upon that part of the plant production which reaches the bottom. However, secondary production and the subsequent trophic level, tertiary production (carnivorous plankton), are consumed in various stages of fish development and must be considered.

Recently the role of starvation—the lack of abundance of the required kind and size of forage at the appropriate time for fish larvae, juveniles, and adults—has become an important consideration in fisheries oceanography studies. Studies of larval fish in the Bering Sea raise many questions because our knowledge of early life stages of fish is extremely fragmentary. Although some information is available on age-one pollock and juvenile yellowfin sole distributions, there are vast gaps in our knowledge of the distributions and movements of these juveniles and those of other commercial species for a year, or in some instances even several years, before they are recruited to the fishery, and we are almost totally ignorant about juveniles of non-commercial species. This has not been a primary concern in managing adult stocks, but if one considers the eastern Bering Sea as a potential area for shelf ranching, a concept approaching reality as a result of the Fisheries Conservation and Management Act (FCMA), an entire spectrum of fisheries and oceanographic information will be required to define this ecosystem.

Although fisheries oceanography studies are not limited to adult commercial fish and their environment, discussions in this section are restricted in this way because studies on physical and biological oceanography, benthos, ice, and other aspects of the Bering Sea ecosystem are presented in other sections of this volume. However, in studies of adult fish the term fisheries oceanography does not mean merely matching climatic or casual marine observations obtained during fishing or other cruises with catch data in an attempt to establish one-to-one relationships; it carries the connotation of controlled studies of relationships among environmental conditions, processes, or events, and species behavior and multispecies interactions. It signifies carefully designed experiments to define relationships or verify suspected associations. Such experiments, with the possible exception of parts of the Processes and Resources of the Bering Sea Shelf (PROBES) program, have not been conducted in the eastern Bering Sea even though a great deal of general information has been obtained from foreign fishing fleets, specifically Japanese and Soviet. Because of management

demands, much of the U.S. research effort in the eastern Bering Sea has been devoted to what is present, rather than why, and what factors can and do cause fluctuations in distribution and abundance of stocks. But this trend is changing slowly as more attention is directed toward ecosystem models.

BACKGROUND

During a discussion of fisheries oceanography in the eastern Bering Sea at Fairbanks, Alaska, in 1974, participants noted the dissimilar spawning and migration patterns of halibut, yellowfin sole, and Pacific salmon and pointed out that PROBES offered an excellent opportunity to conduct multidisciplinary research in relation to or in concert with fisheries (Favorite 1974a). Although an excellent group of investigators joined PROBES, little effort was focused on any fishery but pollock, and only the egg and larval stages have been studied. Five years later, this opportunity to investigate in greater detail these and other patterns of multispecies interactions and resource-environment relations was welcomed.

It might be said that nature abhors an equilibrium—natural forces do not necessarily operate to maintain the status quo, and often man's intervention is all that is required to tilt the system, sometimes drastically. In our haste to evaluate present conditions we should not ignore the past or assume that steady-state balances between primary production and biomass exist. Except for folklore, our early knowledge of fish components stems largely from Cook's voyage in 1778, during which halibut, cod, salmon, and other fishes and mammals were encountered (Munford 1963), but of course relative abundances were unknown. For roughly a quarter of the 19th century, the shelf area south of St. Matthew Island, with the exception of inner Bristol Bay, was considered a major whaling ground; and the area between Nunivak and Unimak islands was still in use in 1875. The exploitation of whales and fur seals in this area and the subsequent exploitation of cod in the same general area in the late 19th and early 20th centuries must have altered several niches and extant transfers of energy between trophic levels, but there are no data to show cause and effect. Did the reduction in whale populations augur well for fish populations because of an increase in availability of planktonic forage or a decrease in consumption of fish? Did the extensive removal of cod by the fishery permit the current massive crab biomass to establish itself? Will the extensive present fishery on pollock result in a change in the feeding habits of fur seals that could trigger changes in the ecosystem?

The information on recent events is better but still fragmentary. The recent sharp decline of halibut, although not necessarily caused by overfishing, is certainly related to the extensive bottom trawling effort during the last decade or two. The apparently large pollock resource may be due to reductions of older age groups by the fishery, thereby reducing the effects of cannibalism, or to the fact that there appears to be a substantial stock seaward of the shelf serving as a reservoir and source of replenishment. Yellowfin sole abundance was severely reduced after the initial fishing on this stock in the 1960's. Herring abundance has always been variable and attempts today to estimate stock size are subject to close scrutiny. And of course for years catches by Japanese salmon motherships have affected salmon returns to various river systems; the forecast estimates of these returns are based largely on counts of downstream migrants supplemented with data from oceanic monitoring.

There is little need to be defensive about the incompleteness of knowledge concerning environmental processes affecting fisheries and stock assessment in the eastern Bering Sea; such studies and evaluations are still being conducted in the North Sea, where fisheries investigations have been carried out independently and cooperatively by many nations for centuries. It was not until the late 18th century that Vancouver showed the eastern side of Bering Strait to be part of North America and provided a map of coastal features. And it was not until the early part of the 19th century that maps and charts indicated that the name Kamchatka Sea had been changed to Bering Sea. Wilimovsky (1966) has summarized early voyages in this area.

Typically exploitation has come before scientific exploration. By the mid-19th century, whaling grounds over the southeastern Bering Sea had largely been abandoned and whalers were entering the Arctic Ocean. At the turn of the century, the cod on Slime and Baird banks north of the Alaska Peninsula were heavily fished and numerous salmon canneries were in operation in Bristol Bay and surrounding areas. Yet it is only today that an understanding of oceanographic conditions and processes is being attained, largely through studies by OCSEAP, PROBES, NWAFC, and other agencies.

Knowledge of bathymetry has also been fragmentary. Although at the end of the 19th Century U.S. Coast and Geodetic Survey and U.S. Bureau of Fisheries charts revealed the existence of the broad shelf and its abrupt termination at the edge of the Bering Sea basin, it has been only within the last decade that a large 300-m depression shown on

Japanese and Soviet charts between the Pribilof Islands and St. Matthew Island was found to be nonexistent (Favorite 1974b). And only within the last few years has some of the complexity of the bathymetry at the shelf edge been discovered as a result of fishing operations.

FISHERIES OCEANOGRAPHY

There are several periods during which fisheries and oceanographic data have been collected with wide-ranging degrees of synopticity and thoroughness. The whalers in the mid-19th century recorded casual environmental data that when combined with data from geographical explorations provided insight into resources and conditions (Dall 1882—see also Ingraham, this volume). At the end of the century the U.S. Bureau of Fisheries conducted studies from aboard the steamer *Albatross*, but after that, fisheries or oceanographic studies were not resumed until the 1930's; except for the World War II period field efforts have markedly accelerated up to the present time.

Steamer Albatross

Rathbun (1894) reported that the *Albatross* explorations were noteworthy in that they constituted an innovation in support of the fishing industry beyond anything ever attempted by any other nation. In a summary of cruises from 1888 to 1892 in the eastern Bering Sea he noted that although three cod banks had been recognized north of the Alaska Peninsula—Slime, Baird, and Kulukak—these banks were not necessarily separate and distinct, except that the westernmost, Slime Bank, was characterized by immense numbers of a large jellyfish, brownish or rusty in color, measuring from 15 to 45 cm across, with long tentacles having great stinging powers. Although not observed upon the sea surface, they appeared abruptly around July 1 in subsurface concentrations which interfered with lowering fishing hooks, using cod trawls, and even raising anchors. Today the general location may be construed as a frontal zone between oceanic and shelf water, but other circumstances are certainly involved in the phenomenon. The largest and best cod during this summer period were taken 11-14 km north of the peninsula; those inshore were of smaller size and inferior quality. Even today there is little specific information concerning the nature or structure of a front along the coast. Rathbun noted also that problems concerning the habits of fur seals on the Pribilof Islands, near the outer edge of the bank,

made it very important that the physical and biological features of the surrounding area be thoroughly studied, and that the distribution of cod was greatly influenced by the movements of capelin, herring, and sand lance but scarcely anything was known regarding the habits of these species on the Alaska coast. However, shortly after the turn of the century the *Albatross* cruises in this area ceased. Only in the last few years have such studies been resumed.

Salmon studies, 1938-41

Except for some tagging studies in the 1920's (e.g., Gilbert 1924) near Port Moller of salmon returning to Bristol Bay streams, there is little evidence of marine research in the area until the 1930's, when the Japanese established a crab fishery and a small operation for bottomfish on the shelf. When it became apparent in 1937 that these operations were to extend to salmon fishing, funds were made available to the U.S. Fish and Wildlife Service (FWS) to conduct investigations of the migration routes and availability of salmon in western Alaska, particularly salmon of the Bristol Bay region. Oceanographic studies were conducted aboard the U.S.C.G.T. *Redwing* in 1938 and salmon fishing studies aboard chartered fishing vessels in 1939; both operations continued until 1941, when the outbreak of World War II interrupted them and delayed the reports of results (Barnaby 1952, Favorite and Pedersen 1959, and Favorite et al. 1961). Although salmon were caught in all areas fished over the southeastern part of the shelf, they were apparently more abundant in the area 54-108 km offshore along the north side of the Alaska Peninsula. The cause of this abundance is still unexplained, although it has been ascertained that seaward-migrating smolts occur inshore of the shoreward migration of adults (Straty 1974). Further, there was a marked shift in dominance from sockeye to chum salmon north of the Pribilof Islands, in an area believed to represent a migration path of the latter to more northern coastal streams. Prophetically, Barnaby noted that there were large populations of bottomfish and shellfish in the area and that it was only a matter of time before exploitation would occur.

Crab studies, 1941

In 1940 Congress approved a special appropriation authorizing the FWS to conduct a king crab study off the coast of Alaska and operations in the Bering Sea were conducted from April to September in 1941 (Fishery Technology Laboratory 1942). Bottom isotherms indicated a tongue of cold water between Cape Newenham and the Pribilof Islands in the direction of Port Moller; sharp gradients along the

north side of the Alaska Peninsula were believed to cause rapid changes in water temperature relatively close inshore, particularly between Amak and Seal Islands, where the greatest concentrations of crabs were found. But since similar conditions did not occur in other productive areas, the significance of this possible relationship could not be evaluated. However, it was pointed out that the incidental catch of edible flatfish was phenomenal. Larger than usual cod were found in offshore waters at depths deeper than usually fished commercially; pollock and yellowfin sole were found in great abundance, and the possibility of a commercial fishery for halibut was indicated. These studies also ceased at the outbreak of the war.

Bottom trawling and oceanography, 1948-49

Crab studies resumed in 1947 aboard the charter vessel *Alaska* (King 1949) in a small area about 75 km northwest of Port Moller. When conditions permitted, air, surface, and bottom temperatures were obtained. The maximum crab catch occurred at a bottom temperature of 3 C and catches were made at temperatures from 1.65 to 7.25 C. Unusually large incidental catches of pollock and yellowfin sole were roughly an order of magnitude larger than cod catches and undoubtedly this prompted the chartered mothership operation (*Pacific Explorer*, with a fleet of 10 fishing vessels) from April to July 1948 in the Amak Island and Black Hills area (Wigutoff and Carlson 1950); but no environmental data are reported.

Perhaps the first fishery cruise that seriously considered oceanographic conditions was the chartered *Deep Sea* from Unimak Pass to Norton Sound during the last week in June and the first week in July 1949. The report of the cruise (Ellson et al. 1950) indicates awareness of previous oceanographic studies (Ratmanoff 1937, Barnes and Thompson 1938, Goodman et al. 1942, and others), and results indicated: there was a positive correlation between depth and bottom temperatures except for the corridor of subzero bottom water extending past St. Matthew Island to the southwestern portion of St. Lawrence Island; bottom water temperatures, more than any other condition, were positively correlated with the abundance of fish taken over the entire area; temperatures at the surface and bottom were several degrees lower than those observed in the previous September from aboard the U.S.F.W.S. *Washington*; summer warming influenced the northward migration of many fish; the cold corridor effectively acted as a faunistic barrier to the movement of many species through the strait between Siberia and St. Lawrence Island; the fauna in this area consisted mainly of eelpouts,

liparids, sculpins, and Tanner crabs; the corridor was believed to extend past St. Matthew Island and the Pribilof Islands into the southeastern Bering Sea, and to cause species entering Bering Sea to be shunted through the Aleutian passes eastward before continuing northward; cod and flatfish in particular were assumed to "lead" along the edges of the cold water barrier, an assumption which would explain the concentrations found in the approaches to Bristol Bay and northward along the warm water channel adjacent to the coast.

These cruises indicate that a number of years ago there was ample evidence of relations between these resources and the environment to justify conducting specific fisheries oceanography studies that would enhance conservation and management practices.

Recent studies

Commercial exploitation of resources, largely by the Japanese, intensified in the 1950's. Scientific studies and conservation measures were enhanced by the formation in 1953 of the International North Pacific Fisheries Commission (INPFC), whose reports, documents, and bulletins are readily accessible and describe Japanese and U.S. investigations, in progress even today, of crab and fish resources. And the NWAFC has conducted an extensive annual survey of resources in the southern part of the area since the early 1970's. But environmental observations are limited and only incidental to fishing effort.

The extensive Soviet Bering Sea Comprehensive Scientific-Commercial Expedition began in 1958; it resulted in numerous compilations of environmental data summarized by Favorite et al. (1976) and fishing reports (Moiseev 1963-72) cited throughout the following chapters. Even though this Soviet activity prompted extensive analyses of existing oceanographic and fisheries data, and environmental observations obtained from aboard fishing vessels contributed to the knowledge of resource-environment relations, such observations were incidental to the fishing effort, fragmentary, and not available through normal channels. Furthermore, no major Soviet oceanographic vessel participated in the investigations.

Recently there has been a trend toward cooperative international fisheries research that becomes more coordinated and successful each year. Emphasis is still, however, largely on fishing at predetermined station locations that provide assessments of resource abundance; little funding, personnel, and vessel time are given to specific fisheries oceanography studies that may provide the key to more efficient and effective resource assessment and improved management and conservation.

SECTION SUMMARY

Many of the physical and biological characteristics of the eastern Bering Sea are directly related to fisheries oceanography. First, it has an unusually broad, relatively flat continental shelf the slope of which abruptly increases mainly near the 150-m isobath. It is relatively isolated from the currents of the North Pacific basin by the Alaska Peninsula, the Bering Sea basin by the abrupt shelf edge, and the Arctic basin by the narrow and shallow Bering Strait. It has three major embayments—Bristol Bay, Norton Sound, and the Gulf of Anadyr. It is basically a subarctic environment with ice cover present and expanding seaward from October to a maximum in April, and largely absent by June. Although surface and coastal water temperatures in summer are not unlike those as far south as the Washington coast, the -1.8°C water temperatures under the ice in winter and near the bottom at mid-shelf during summer identify the Bering Sea shelf regime. High river runoff in late spring markedly alters coastal water properties and flow, and oceanic perturbations greatly affect flow at the outer portion of the shelf; whereas conditions at mid-shelf are largely influenced by tidal currents. Warming and the formation of a surface layer of ice melt in summer result in positive stability; during the winter, ice formation produces slightly negative stability, which causes water overturn.

The sluggish circulation and nutrient replenishment as a result of winter overturn are conducive to high production and standing stocks in spring and summer; the ice in winter causes extensive fish migrations to the deeper, warmer water at the shelf edge and upper slope. Pollock are the dominant semidemersal species, yellowfin sole the dominant demersal species, and herring a dominant pelagic species although the anadromous salmon are abundant from June to September.

Although the largely discipline-oriented approaches by individual OCSEAP research units and the generally independent field surveys make it difficult to make multispecies or interdisciplinary assessments, the opportunity to bring the fragmentary information on the various life stages of major biological components up to date, to assess dynamic aspects of the ecosystem, and to extend the available knowledge through the use of simulation techniques presented a great challenge; and that is what we as a group have attempted to accomplish—in spite of frustrations associated with the lack of continuity of data in space and time, the multiplicity of sampling techniques, and large gaps in basic knowledge. We have had to be selective in our approach and limit discussions of

resource-environment relations to several commercial fish which obviously prey on zooplankton, other fish, and benthos, and are preyed upon by fish, birds, and mammals. Even though such interactions are poorly documented, numerical assessments have been made using simulation techniques.

First, background information on environmental conditions (Ingraham) and distributions of ichthyoplankton (Waldron) is presented. Next, distributions, abundance, migrations, spawning activities, and environmental relations of five selected species are summarized: Pacific halibut (*Hippoglossus stenolepis*), which spawn seaward of the shelf in winter and migrate shoreward in spring (Best); Pacific herring (*Clupea harengus pallasii*), some of which migrate inshore across the shelf in spring and spawn along the coast from April to July, and others winter inshore (Wespestad and Barton); walleye pollock (*Theragra chalcogramma*), which spawn perhaps ubiquitously in subarctic waters but certainly in large numbers inshore in spring along the shelf edge (Smith); yellow-fin sole (*Limanda aspera*), which migrate inshore across the shelf in late spring and early summer and spawn in the southeastern part of the shelf (Bakkala); and salmon (*Oncorhynchus* spp.), which as adults migrate shoreward across the shelf to various coastal areas in late spring and early summer to spawn and, as juveniles, migrate seaward across the shelf to oceanic areas in spring and summer (Straty). Then, having provided a general, descriptive assessment of temporal and spatial variations in fish distributions, migrations, and spawning locations in relation to environmental conditions or events (Favorite and Laevastu), we evaluate the total biomass of ecosystem components and their interactions, using simulation techniques (Laevastu and Favorite).

After having perused the following chapters, one cannot help feeling that OCSEAP studies, rather than having completed baseline data acquisitions for the eastern Bering Sea shelf and achieved basic or preliminary understanding of processes, have actually only permitted us to define the nature of the tasks that have to be accomplished if we are to obtain a sufficient understanding of the eastern Bering Sea ecosystem to assess the impact of oil exploration and exploitation and protect the extant living marine resources.

Up to this point little new information has been obtained and previous knowledge has been only marginally extended. One talks about warm and cold years, but nothing is known about what occurs under the ice. One attempts to measure primary and secondary production without concern about patchiness, predation, and reproduction, or evidence

of biomass balances. One conducts casual studies of ichthyoplankton without regard to the limitations in our knowledge of their identification or location and never pursues organisms much beyond the egg stages. One knows little or nothing about the distributions or migrations of juveniles for the years before they enter the fishery as adults. One notes that cold conditions appear to delay shoreward migration of adults but also that in some instances migrations of more northerly stocks do not appear to be influenced at all by equivalent temperatures. And, in most cases, knowledge of year-round distributions and migrations of adult fishes and specific causes for aggregations or movements is still severely limited.

It is easy to say that the ocean is a turbulent regime, that each year is different, and that only gross assessment can be made and only general relations ascertained. However, there is an apparent order to the Bering Sea ecosystem that implies in many instances some very specific fish responses that we do not understand; only dedicated, multidisciplinary research (fisheries oceanography) will provide the answers.

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Shelf Environment

29

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ABSTRACT

Monthly mean environmental conditions of ice, temperature, runoff, and salinity for winter (January-March), May, July, and September are presented for the eastern Bering Sea shelf area. Mean geostrophic flow (0/50 db) in summer (July) reflects the northwesterly flow at the shelf edge and a westward flow from the Yukon River area seaward south of St. Lawrence Island. Tidal currents as derived from a hydrodynamical-numerical model reflect a NW/SE flow over the outer shelf and NE/SW flow north of the Alaska Peninsula, between Nunivak and St. Matthew Islands, and in the southern Gulf of Anadyr. Although data are fragmentary, environmental conditions over the shelf are highly variable and examples of N-S surface wind components (1946-75), ice cover (1976-78), and bottom temperatures (1969, 1973, 1976, and 1977) reflect departures from mean conditions.

INTRODUCTION

The eastern Bering Sea shelf is characterized by several features that have pronounced effects on its hydroclimate. Sloping gently westward over 500 km, the shelf terminates abruptly as the continental slope drops somewhat precipitously into the Aleutian basin, resulting in a coastal-oceanic water interface that largely divides the surface area of the Bering Sea in half. The shelf is isolated from the direct effects of circulation in the Pacific Ocean by the Alaska Peninsula and sheltered from the Arctic Ocean by the narrow, shallow Bering Strait.

Seasonal advance and retreat of ice cover in winter and spring and the extensive discharge of runoff in spring and summer dominate conditions and processes in the hydroclimate. Since physical processes are discussed in another section of this book, we are concerned here primarily with conditions over the entire shelf area, insofar as data are available. For increased perspective on recent studies I will make brief reference to the considerable knowledge gained from early historical studies, present mean conditions for January-March, May, July, and September, and discuss variations in ice, winds, and bottom temperatures.

Although even today the Bering Sea is considered a remote and hostile area, it must have been considerably more treacherous and forbidding to early explorers and particularly whalers with limited resources and technology. The loss of 33 whaling vessels north of Bering Strait in 1871, only four years after the United States acquired Russian America, resulted in environmental studies in the northern shelf area (e.g., a survey of Bering Strait from U.S. Schooner *Yukon* in 1880) and an assessment of logs and reports from whaling and other vessels crossing this area (Dall 1882); considerable information was derived.

First, on the subject of ice and surface temperatures: (1) the southern limit of ice was shown first to extend to 56°N in the central Bering Sea and well below the Pribilof Islands in the shelf area; (2) in cold years ice remains around St. Paul Island until late May; (3) ice opens to the west of St. Lawrence Island first, for the sea is frequently navigable there and north of the island into Norton Sound at a time in the season when the passage between Nunivak Island and St. Lawrence Island is still blocked with decaying ice; (4) water opens first where the southerly cold set away from the ice is strongest; (5) the extent of ice in different years varies and depends on winds, with prolonged winds from the north bringing southward loose flow ice which is prevented by the formation of new ice from returning northward when the winds change; (6) the water retains nearly its normal temperature to within a very short distance from a large field of ice; (7) the Yukon discharge during summer months gives rise to definite local currents; (8) in summer at depths of 18 m (10 ftm) a higher temperature is found than in adjacent deeper water; (9) surface temperatures in Norton Sound obtained at anchor on 7 July over a 24-hour period varied from 6.7 to 12.2 C, a range of 5.5 C. Recent

cruises and satellite imagery have largely borne out these early observations.

Second, on the subject of surface currents it was recognized that: (1) discharges from numerous large rivers and the westerly and northerly marching tide create a distinct set in the vicinity of their mouths more or less directed by local winds; (2) the Kuskokwim and Bristol Bay rivers have a southwesterly or southeasterly set depending on prevailing wind and tide; (3) discharge from the Yukon River in calm weather proceeds in a northwesterly direction, but under northerly winds or ebb tide, or both, part or all of it passes to the south of St. Lawrence Island; (4) away from the influence of the rivers the current is influenced a great deal by the direction and force of the wind; (5) north of Nunivak Island ($60^{\circ}41'N$, $166^{\circ}04'W$) the U.S.S. *Corwin* reported NW/SE tidal currents of 0.5-2 kn (~ 25 -100 cm/sec) while in the ice; (6) at St. George Island a steady current of 1-2 kn (~ 50 -100 cm/sec) from the west is sometimes observed for several days; (7) currents across Bering Strait (0705-1415 hrs, 5 Sept. 1980 from aboard U.S.S. *Yukon* under fresh northwesterly wind, transit W to E) varied as follows: E $21.8^{\circ}S$, 3.4 kn; N $24.3^{\circ}W$, 1.5 kn; W $26.3^{\circ}S$, 3.8 kn; S $1.3^{\circ}E$, 2.5 kn; W $31.0^{\circ}N$, 2.2 kn. Numerous reports of drifts of whalers under various conditions of ice, winds, and sea were presented for analysis and provide considerable insight not only into tidal flow but also into general drift. All this information reflects the complexity in flow and the difficulty even today of evaluating flow from a limited number of current meters.

Third, although subsurface data were sparse because of limited instrumentation, there were reports of conditions across Bering Strait and of environmental divisions over the shelf of considerable value. The U.S.S. *Yukon* hydrothermal cross section of Bering Strait in September 1880 clearly indicated: the 6 C temperature difference across the strait—vertically isothermal conditions of 2.6-3.3 C (36.7-38 F) in the western third, the temperature gradient (3.3-7.2 C) regime in the central third, and the slight stratification of water of 7.8-9 C (46-48 F) in the eastern third with maximum values near the surface, offshore of the eastern side of the strait. However, in spite of the temperature differences implying that opposing currents exist in the strait, the fact that strong northerly and southerly currents had been experienced on both sides of the strait by various navigators suggested that there was not sufficient evidence to report that a southerly cold current occurs on the west side of the strait, and a warm northern current on the east; but it was recognized that the northerly flow on the east side of the

strait nearshore was composed largely of river runoff. In addition, on the basis of temperature data only, the Bering Sea was separated into three divisions: (A) the shallows (e.g., Bristol Bay, Norton Sound, etc.) and the area seaward to the 46 m (25 ftm) contour; (B) the moderate depths, between the 46 m (25 ftm) and the 137 m (75 ftm) contours, from Bering Strait to the Alaska peninsula; and (C) the deep waters to the south and west. Attempts to improve on this scheme continue even today but no significant deviation from this early assessment has been accepted. Favorite (1974), prior to OCSEAP and PROBES studies, distinguished four subdomains on the basis of temperature-salinity relations: Gulf of Anadyr, West Alaska Coast, Mid-Shelf, and Shelf Edge (Fig. 29-1). Subsequent PROBES studies have focused on the frontal zones between these subdomains only in the southern part of the shelf (e.g., Coachman and Charnell 1979).

Although most investigators are familiar with the University of Washington oceanographic studies of the 1930's (e.g., Barnes and Thompson 1938) and those by the U.S. Bureau of Fisheries in 1938-41 (e.g., Favorite and Pedersen 1959; Favorite et al. 1961), few are aware of the studies by the U.S. Naval Electronics Laboratory in 1949, because portions of them were classified. Some results are available (Saur et al. 1952) that describe general summer conditions, and temperature and salinity relations were used to identify seven water masses in the eastern Bering and Chukchi Seas.

Oceanographic studies conducted from 1953 to the

Figure 29-1. Shelf sub-domains ascertained by T-S relations in $1^{\circ} \times 1^{\circ}$ quadrangles.

present time during fisheries investigations under the aegis of the International North Pacific Fisheries Commission (INPFC) and other studies during this period have been summarized by Dodimead et al. (1963) and Favorite et al. (1976); and recent OCSEAP and PROBES studies are summarized in Chapters 1-8 of this book.

MONTHLY MEAN CONDITIONS

Ice information presented is based on data from Potocsky (1975) and recent satellite imagery; temperature and salinity data are based on the current (June 1979) NODC geofile (including OCSEAP data) and Japanese and NWAFC fisheries cruises. These last data have been computer-processed into monthly means by $\frac{1}{2}^\circ$ (lat.) and 1° (long.) quadrangles, and anomalies are derived from these means; however, only summaries of conditions for winter (January-March), May, July, and September are presented and discussed. There are several limitations to the presentations of mean conditions in this area: first, data are not sufficient for summaries by $\frac{1}{2} \times 1^\circ$ quadrangles (nor even $2 \times 2^\circ$), and thus there are numerous gaps in each month and only limited data from October to March. Second, equivalent data were not acquired in warm and cold years; thus in some instances the values represent mean conditions, in others (in warm or cold years) possibly extremely anomalous conditions. (The numbers of observations and years are not presented here, but are available at NWAFC.) Third, the paucity of data can also result in isolated anomalous values and tongue-like protrusions that may merely reflect a preponderance of data from a single year rather than a significant environmental feature. Nevertheless, the data reflect a high degree of continuity of conditions throughout the area that cannot be obtained in any other manner.

Ice

In the northern Bering Sea, monthly mean air temperatures vary from 10 C in summer to -20 C in winter and extreme conditions extend this range over 10 C higher in summer and lower in winter. Temperatures are mostly below 0 C from October to May and thus sea ice plays a dominant role in the ocean environment. Freezing occurs at about -1.8 C in this area, and the formation of ice limits winter cooling of sea water to this temperature. The actual nature of the ice edge in any one location—compact or spread out—is also influenced by advection of ice due to wind drift.

Much of the increasing solar radiation in spring is consumed in the melting of ice and the seasonal

increase of sea surface temperatures is delayed. However, when ice is no longer present, dilution and abrupt warming at the sea surface result in a shallow, stable surface layer which retains most of the incoming radiation and permits nearly subtropical conditions to occur, particularly in the inshore areas.

The area is free of ice for only a short period, usually from June or July to August or September. Usually by September ice begins to form near Cape Dezhnev and along the western shore of Bering Strait; by October it extends into the Gulf of Anadyr, across Bering Strait, and even into the eastern part of Norton Sound. Mean conditions (Fig. 29-2) indicate a progressive southwesterly advance through March or April to the edge of the continental shelf, but a rapid retreat from May to June, from north of the Pribilof Islands to Bering Strait. Potocsky (1975) presents an excellent summary of mean ice conditions and indicates wide deviations from the monthly means. (See also the ice section of this volume.)

Surface and bottom temperature

Mean surface temperatures primarily reflect the annual cycle of insolation; secondary effects are observed in nearshore areas affected by river runoff and reduced cloud cover, and in areas of ice cover where spring warming is delayed until the ice is melted. Winter (January-March) surface temperatures of less than -1.5 C (Fig. 29-3) reflect the approximate extent of ice cover from Bristol Bay to Cape Navarin. The warmest water, 2 C, is offshore in the ice-free area, and intermediate temperatures of 1-2 C

Figure 29-2. Monthly mean location of ice edge.

Figure 29-3. Long-term mean sea surface temperature (C) for January to March, May, July, and September.

are found just south of the ice on the southern portion of the shelf. By May, temperatures over the southern shelf have increased to 3-5 C, but over most of the shelf relatively low temperatures between -1 and 1 C prevail, marking the initial stage of the heating cycle after the ice melt. Surface temperatures for July show that most of the area has warmed rather quickly to 6-8 C and inshore temperatures have reached a maximum except in the extreme north, with Bering Strait temperatures near 5-6 C and negative temperatures still present on the east coast of Siberia north of the strait. Local warm spots

are seen with temperatures of 12 C in Anadyr and Bristol Bays, and in excess of 16 C in Norton Sound. August is the warmest month for surface water temperatures; the majority of the area reaches 8-10 C, thus reducing the contrast between the offshore areas and the coastal areas, where temperatures are nearly the same as in July. A dramatic change does occur, however, in Bering Strait with the apparent southward movement of cold east Siberian coastal water and the apparent northward movement of the warm Alaskan coastal water, producing a pronounced gradient of nearly 10 C between the cold western and

warm eastern sides of the strait. By September the effects of the beginning of the cooling cycle are evident; most of the area temperatures are between 7 and 9 C and coastal temperatures have decreased to about 10 C or less. The gradient across Bering Strait has decreased to about 6 C.

Bottom temperatures result primarily from the processes of vertical mixing and diffusion and to a lesser extent from the processes of horizontal mixing and advection. Mean temperatures lower than -1.5°C in winter (January-March) were distributed over most of the shelf, extending seaward as far southward as mid-Bristol Bay and as far offshore as St. Matthew

Island, but close to shore in Anadyr Bay (Fig. 29-4). Surface cooling, the formation of ice, and overturn have produced a vertically isothermal water column which extends to a depth of at least 50 m and sometimes as much as 75 m. Warmer water of 2-4 C is found at depth near the shelf break associated with the general cyclonic (counterclockwise) circulation over the Bering Sea basin and the northward extension of Alaskan Stream water from the North Pacific Ocean. Thus, isotherms tend to be roughly parallel to the bathymetric contours. The presence of 0-1 C water just north of the Alaska Peninsula is probably correct in the mean sense, but the serpentine nature

Figure 29-4. Long-term mean bottom temperature (C) for January to March, May, July and September.

of the contours may be misleading due to the absence of data during extreme winters when this area may be less than -1°C , or mild winters when temperatures are probably near $1-2^{\circ}\text{C}$; for this reason, the presence of a northeastward flow which is implied if one interprets the contours as warm advection from the southwest cannot be substantiated. Another area that appears to be influenced by some cross-shelf, warm advection, however, is Anadyr Bay.

By May there has been little change in the bottom temperature distribution in the outer shelf area between the 100 and 200-m isobaths, but the shallow area from inside the 50-m isobath to the Alaskan coastline, from Bristol Bay to the mouth of the Yukon River, has warmed from the winter condition of less than -1.5°C to between 0 and 2°C , with highest values occurring closest to shore. Therefore, the dominant feature is the area of remnant negative temperatures which appears as a large southeastward tongue located about mid-shelf. Apparently the dominant process in the warming of coastal bottom water at this time is downward mixing of surface-warmed water rather than horizontal warm advection from the south. The negative isotherms associated with the tongue appear discontinuous and in isolated areas; again, this is probably the effect of inadequate temporal distribution of data.

Because the best data coverage is for midsummer, mean bottom temperatures for July show the most complete picture of shelf temperatures. The coastal area has warmed to more than 10°C from Bristol Bay to Norton Sound, thus greatly accentuating the inshore edge of the cold tongue, where a large temperature gradient ($10-12^{\circ}\text{C}$) occurs. Temperatures at the seaward side of the tongue remain about the same north of the Pribilof Islands. The cold core of the tongue appears to warm slightly as the portion south of 58°N is now more clearly defined by the 1°C isotherm; whereas to the north small remnant patches of $<-1.5^{\circ}\text{C}$ water still occur. An increase of about 1°C is seen in a separate tongue advecting warm ($>4^{\circ}\text{C}$) water northwestward from Unimak Island. Similar to surface temperatures, bottom temperatures across Bering Strait reflect about a 4°C west-to-east gradient.

By September the warm coastal water ($8-10^{\circ}\text{C}$) reaches its greatest seaward extent, thus sharpening the already high gradient on the eastern side of the cold tongue and shifting it slightly seaward to about 100 km offshore. A division appears to develop in the cold core near 58°N as $2-3^{\circ}\text{C}$ water influenced by the westward movement of the warmer coastal water appears to separate the tongue into two areas of temperatures lower than 1°C , to the north and south. The reappearance of some negative temperatures in

the southern core reflects additional September data in cold years, because July temperatures are higher ($0-1^{\circ}\text{C}$), and it is unlikely that the apparently colder water in September was advected from the north across the intermediate warm area. In Norton Sound bottom temperatures increase to $6-8^{\circ}\text{C}$. And on the northern edge of the cold core, a tongue of 2°C water intruding from the west replaces the negative temperatures of July in Anadyr Bay.

River runoff

Although data on river runoff are sparse, general characteristics of the area may be deduced from data for the Yukon River, the largest individual source of fresh water. The Yukon River drains 45 percent, the Anadyr River 20 percent, the Kuskokwim 12 percent, the Nushagak 3 percent, and the remaining rivers 16 percent of the total land area draining into the Bering Sea. Because of regional differences in precipitation, however, it may be misleading by as much as a factor of two to estimate the magnitude of discharge from the major rivers by comparing their drainage areas alone. The only data for an extended period of time are for the Yukon River at Ruby where 22 years of monthly mean discharge data are available. Long-term monthly averages (Fig. 29-5) indicate a pronounced seasonal pattern of runoff; low during the freezing period, decreasing from $2 \times 10^3 \text{ m}^3/\text{sec}$ in November to $1 \times 10^3 \text{ m}^3/\text{sec}$ in April, it increases rapidly to a peak of $13.5 \times 10^3 \text{ m}^3/\text{sec}$ in June and then steadily decreases to low levels by November as the freezing period begins.

Some examples of variability are evident in the last four years, which are characterized by alternating high and low values. Beginning in 1975, above-normal spring and summer runoff was followed by low runoff in spring and summer of 1976. This pattern was repeated with above-normal runoff in spring 1977 and below-normal values during spring and summer 1978. The lowest values in the last four years occurred in 1978, and a shift in timing of the peak discharge occurred from June to July. This shift also occurred in two other low-discharge years, 1960 and 1963. For three consecutive summers (1976, 1977, and 1978) during the period after the annual peak discharge, runoff has been below normal.

Surface and bottom salinity

Mean surface salinities increase from November to April by the processes of salt exclusion during freezing and vertical mixing and then decrease from May to October by precipitation and river runoff, which delivers a large pulse of fresh water to the surface layer. Maximum values ($>33.3^{\circ}\text{oo}$) occurring in

Figure 29-5. Monthly mean Yukon River runoff (m^3/sec) at Ruby 1975-78 and long-term (22-year) mean.

northern Anadyr Bay and north of St. Lawrence Island (Fig. 29-6) are believed to be local results of the freezing process inasmuch as they are separated by large areas of lower salinity. Values lower than $32.0^\circ/\text{oo}$ occupy the majority of the shelf area south of St. Lawrence Island and out to about the 100 m isobath. By May several changes are evident. A decrease of $1\text{--}2^\circ/\text{oo}$ is evident in inner Bristol Bay, in Norton Sound, and north of St. Lawrence Island, and a long tongue of values less than $31.0^\circ/\text{oo}$ extends eastward from the Yukon River mouth to just south of St. Lawrence Island. By July ice is no longer present and the river runoff has resulted in a pronounced dilution of surface water along the west coast of Alaska and in Anadyr Bay; a minimum value of less than $16.0^\circ/\text{oo}$ occurs in Norton Sound. Despite this marked dilution inshore and lesser changes offshore (evident in the westerly displacement of the $33.0^\circ/\text{oo}$ isohaline—nearly 200 km), the $32.0^\circ/\text{oo}$ isohaline appears to be in a consistent position roughly paralleling the 100-m isobath along

an approximate line from Unimak Island to Cape Navarin. By September the effect of further dilution seaward of the shelf is evident as very little water of greater than $33.0^\circ/\text{oo}$ remains over the basin, but conditions in the shelf area are similar to those of July except for slightly higher salinities near the Yukon River mouth and in Norton Sound, probably the result of mixing and decreased runoff.

Mean bottom salinities below 50 m generally increase with depth. In January-March values of bottom and surface salinity are largely the same in water less than 50 m deep—essentially isohaline conditions (Fig. 29-7). In water deeper than 50 m, however, stratification appears to develop, as shown by the displacement of both the 32.0 and 33.0 bottom isohalines about 100 km farther shoreward (northeastward) than the corresponding surface isohalines. This salinity increase with depth allows the higher temperatures at these depths to exist in a vertically stable density field. There is little change in bottom salinities in May except for the presence of some slight dilution to less than $31.0^\circ/\text{oo}$ in inshore Bristol Bay and south of St. Lawrence Island. By July, however, strong dilution has occurred in Norton Sound with values of less than $24.0^\circ/\text{oo}$ and to a lesser extent in Bristol Bay ($< 28.0^\circ/\text{oo}$). A notable feature is the absence of dilution in Anadyr Bay at the bottom despite marked dilution at the surface; this condition persists through September as the area with salinities greater than $33.0^\circ/\text{oo}$ continues to enlarge. Also in September dilute water extends farther offshore in Norton Sound and bottom salinities increase slightly in inshore Norton Sound and Bristol Bay, probably because of horizontal mixing.

Water masses as reflected by surface and bottom temperatures

There have been numerous attempts to describe water masses in the Bering Sea using the classical method of T-S (temperature vs. salinity) diagrams to recognize groups of data representing water of similar characteristics, which are put into envelopes based on dominant processes and/or the resulting water structure. Recent studies (e.g., Takenouti and Ohtani 1974) have considerably more detail than the early ones, and OCSEAP studies have refined considerably our knowledge of water masses and their interactions over the southern shelf offshore from Bristol Bay (Kinder 1977, Kinder et al. 1978); however, four major distinct environments stand out—Gulf of Anadyr, West Alaska Coast, Mid-Shelf, and Shelf Edge (Favorite 1974).

Although salinity characteristics influence the water mass definitions in the Gulf of Anadyr (where

Figure 29-6. Long-term mean sea-surface salinity (‰) for January to March, May, July and September.

high salinities are associated with low temperatures due to the salt exclusion process) and also near the shelf edge area (from the influence of Bering Sea basin water), the most striking feature of the shelf environment is the temperature regime, particularly in the mid-shelf and coastal waters. The coastal water mass is clearly distinct from the others because of its relatively constant vertically homogeneous nature, whereas marked stratification occurs in summer at mid-shelf. Thus, to illustrate the extent and seasonal changes in mean water mass conditions, it

is expedient to look at temperature differences between surface and bottom.

During January-March, mean temperature differences (surface minus bottom) are near zero over most of the shallow shelf as ice cover and deep vertical mixing result in isothermal conditions (Fig. 29-8). The most significant feature is the oblong area of relatively large negative differences (-2 to -4 °C) which occurs all along the outer shelf, where seasonally cold surface temperatures override a deep, warm, slightly more saline water mass. By May there

is little change except west of St. Lawrence Island where two extreme values of -2°C and 2°C indicate local conditions at the beginning of spring warming after completion of ice melting. Dramatic changes are seen by July and continue through September as the dominant summer feature, a difference of $6-8^{\circ}\text{C}$, develops over the remnant tongue of cold mid-shelf bottom water. The eastern side of this area is clearly defined by the sharp decrease in temperature difference, which approaches zero in coastal water; the western side, near the outer shelf, is less well defined since the difference is due to the only slightly warmer bottom temperatures.

Flow

Although the most reliable data on flow come from direct current measurements, few data have been taken and only in local areas of individual interest such as Bering Strait and Bristol Bay (Coachman and Charnell 1979), so that indirect methods are still required to infer net flow or mean circulation; these methods have proved effective in gaining knowledge of the general flow in the climatological sense which concerns us here. One must keep in mind that since tidal components on the order of 50 cm/sec occur daily which reverse either diurnally or semidiurnally, any estimate of mean flow

Figure 29-7. Long-term mean bottom salinity ($^{\circ}/_{\infty}$) for January to March, May, July, and September.

Figure 29-8. Long-term mean surface minus bottom temperatures (°C) for January to March, May, July, and September.

will probably be at best about one order of magnitude less than the maximum of the instantaneous flow. An NWAFC tidal model study (Hastings 1975) indicated that the reversals of flow are largely NE/SW in inner Bristol Bay, between Nunivak and St. Matthew islands, and between St. Matthew Island and the Gulf of Anadyr; and NW/SE over much of the re-

In the absence of direct measurements, the geostrophic approximation is widely used to calculate mean currents from the vertical distributions of temperature and salinity relative to an assumed level of no motion, usually below 1,000 m depth. This method has been used to establish the existence of northwestward mean flow of the Transverse Current (Favorite et al. 1976) seaward of the shelf edge, but is invalid in shallow water over the shelf except for showing anomalies of specific volume from a geometric level. The mean data for July did cover sufficient area to warrant examination, and contours

Figure 29-9. Tidal currents (cm/sec) calculated from numerical tidal model for (a) four hours before high tide, and (b) four hours before low tide at the eastern boundary (from Hastings 1975).

of anomaly of dynamic height computed from the surface relative to 50 db (0/50 db) suggest several major features of flow (Fig. 29-10). Northwestward flow is indicated over the basin between Unimak Island and Cape Navarin and a component veers to the right around Anadyr Bay, eastward to St. Lawrence Island, and finally northward through Bering Strait. There is a suggestion of a weak westward cross-shelf flow just north of St. Matthew Island and also a northeastward flow into Bristol Bay along the north coast of the Alaska Peninsula, with a westward meandering return flow out of northern Bristol Bay. Although the action of surface winds could result in markedly different flows in the upper and lower portions of this 0-50 m layer, which are well isolated in summer by thermal and haline stratification, these results are interesting and should form the basis for further investigations.

VARIABILITY

McLain and Favorite (1976) have presented monthly anomalies of air and sea surface temperature in the vicinity of the Pribilof Islands (57°N , 170°W) from 1967 to 1975, which reflect not only a long-term downward trend (which reversed itself in 1977) but also considerable monthly and annual variability. Mean monthly north and south geostrophic wind components computed for 1946-75

for the grid point 57°N , 170°W (Fig. 29-11) indicating potential periods of cooling (and ice advance in winter and spring) and warming (and ice retreat in winter and spring) reflect the extreme variability of meteorological conditions; and the 12-month running mean indicates an approximate three-year cycle and a long-term trend of increasingly dominant northerly

Figure 29-10. Long-term mean anomaly of dynamic height (0/50 db) for July.

Figure 29-11. Monthly mean wind (m/sec) components at 57°N, 170°W, 1946-75. (Positive values indicate northward component, negative values indicate southward component.) Twelve-month running mean showing annual mean trends.

winds (cold conditions). Kihara (1971) has reported the extreme variability of the extent of Alaskan Stream Extension water over the southern part of the outer shelf. Although polar waves considerably alter air temperature, wind, and precipitation, annual variability in shelf conditions is most easily shown by the extent of ice cover and by bottom temperatures.

Ice

In recent years (1976-78) the extent of ice cover for November (when ice first extends seaward) was minimal in 1976 and maximal in 1977; whereas for April (usually the month of greatest ice cover), the maximum extent occurred in 1976 and the ice edge has been farther north each succeeding year, with the greatest changes occurring on the eastern side of the shelf (Fig. 29-12). Thus, there is no correlation between spring conditions and subsequent fall conditions (i.e., summer conditions can obliterate cold trends established in the spring).

Furthermore, although the normal southerly winds in spring result in a compact ice edge and hasten its

northerly retreat, northerly winds can cause anomalous conditions. In April 1976 an unusual and large southward displacement of part of the ice edge in the Bristol Bay area resulted in ice immediately north of the Alaska Peninsula which forced vessels out of the area (Northwest Fisheries Center 1976).¹ This displacement not only caused water in the open area in the northern Bristol Bay to be subjected to early insolation (and thus warming), but because spring warming at the surface had already begun, considerably altered the temperature structure of the water in the area into which the ice was advected, resulting in extremely anomalous conditions in spring 1976.

Bottom temperatures, 1973 and 1977

Because of the scarcity and fragmentary nature of bottom temperature data, I have chosen to illustrate

¹ Northwest Fisheries Center, 1976, Ice conditions in the eastern Bering Sea, April 1976. Northwest Fisheries Center Monthly Report, April 1976, 7-9. Northwest and Alaska Fish. Cent., Nat. Mar. Fish. Serv., NOAA, Seattle, Wash.

Figure 29-12. Mid-month position of the ice edge, April 1976-79 and November 1976-78.

variability in the months of July 1973 and July 1977, for which time data were available for Norton Sound and Bristol Bay, even though it is known that extremes of temperature occurred in other years (warm in 1968-69 and 1977-79, cold in 1971 and 1976). Bottom conditions reflect general surface conditions (warm or cold, not specific temperatures); the reverse is not true. These data indicate that in summer a general shelf categorization of warm or cold years may be unreliable for determining bottom conditions—in July 1973 positive anomalies of 3 C occurred in Norton Sound and negative anomalies of 2 C in Bristol Bay. However, in July 1977 negative anomalies in excess of 4 C occurred in Norton Sound and positive anomalies of 2 C in Bristol Bay (Fig. 29-13). Thus, anomalies of bottom temperature of different signs and magnitudes may occur over relatively short space scales within the shelf area. It is apparent that before one can consider movements of fish throughout the shelf, one needs to know the environment over the entire shelf.

Figure 29-13. Anomalies of bottom temperature (C) from the long-term mean, (a) July 1973 and (b) July 1977.

*Bottom temperatures,
June and July 1969 and 1976*

One further example of variability is given by comparing bottom temperatures in June and July 1969 with those of June and July 1976 (Fig. 29-14). The temperature difference of warm conditions in 1969 (also in 1978) from cold conditions in 1976 was as much as 5 C. These conditions are shown in later chapters to affect movements of fish but they must also have significant effects on the survival, reproduc-

tion, and movements of the benthos and other creatures of the shelf.

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Figure 29-14. Anomalies of bottom temperature (C) from the long-term mean, June and July 1969, and June and July 1976.

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Ichthyoplankton

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ABSTRACT

Ichthyoplankton studies in the Bering Sea have been conducted since 1955 by Japanese, Soviet, and United States biologists. Comparison and integration of the results of the various surveys is difficult because surveys were conducted during different seasons, different nets and types of tows were used to collect the samples, and different areas were surveyed. Sampling effort was unevenly distributed seasonally with only 1 percent expended during winter, about 9 percent during fall, 35 percent during spring, and 55 percent during summer.

Of the approximately 300 species of fish that occur as adults in the Bering Sea, eggs and/or larvae of about 270 species, divided among 137 genera in 34 families, might be expected to be present in plankton samples. Plankton collections made since 1955 contained 60 species divided among 52 genera in 24 families. Families that occurred most frequently in the 26 collections studied were Cottidae (26), Gadidae (23), Hexagrammidae (23), and Stichaeidae (23). Because of differences in seasonal and areal coverage, it is difficult to determine which larvae were most abundant. For collections made during spring between the Aleutian Islands and about 60°N and centered over the continental slope, larvae of pollock (*Theragra chalcogramma*) were much more abundant than any other species or genus.

An adequate knowledge of the ichthyoplankton of the Bering Sea can be gained only by a comprehensive survey, possibly a cooperative effort by Japanese, Soviet, and U.S. vessels and scientists, of at least a year's duration.

INTRODUCTION

Ichthyoplankton data for the eastern Bering Sea are not so extensive as for some other areas (e.g., off the California coast) and most of the work has been done since 1955. Between 1955 and 1975 most of the collections were made on cruises of the Japanese fisheries training ship *Oshoro Maru* and on a number of Soviet expeditions to the Bering Sea. Before 1975 there were only three reports compiled by U.S. biologists concerning ichthyoplankton in the Bering Sea. This chapter summarizes all available data east of 174°E and points out inadequacies in the information. It should be obvious that although damage from oil spills in local areas may cause infinitesimal losses to generally ubiquitous forage, not only the

loss of forage in extensive spawning areas but the damage to ichthyoplankton itself are major problems. The assessment of damage will depend on knowledge of the temporal and spatial distribution of fish eggs and larvae.

Beginning in 1975 the United States sponsored two programs, OCSEAP and PROBES, which included studies of ichthyoplankton of the Bering Sea, especially over the outer continental shelf. From the inception of these two programs to the summer of 1978, 10 cruises studied ichthyoplankton as a primary or secondary objective.

Cruises in the Bering Sea since 1955 from which came reports dealing with general ichthyoplankton or a single species of ichthyoplankton are listed in Table 30-1. Information concerning some of the cruises made in 1978 is in the form of preliminary cruise reports, and reports of cruises of the *Oshoro Maru* made since 1969 indicate only the number of stations and type of collections made. Some of the reports include only taxa in the total collections, others taxa by station; a few note only the distribution and abundance of a single species. There are other reports of collections of zooplankton in the Bering Sea, some of which include a general category of fish eggs or larvae, but such reports are not included in the table or elsewhere in this summary.

PROBLEMS OF COMPARISONS AMONG CRUISES

It has been necessary to use sources of information with different degrees of completeness and accuracy in the preparation of this summary. Information concerning Japanese cruises aboard the *Oshoro Maru* includes the location of stations at which larvae were caught but not stations at which eggs and larvae were not caught (i.e., non-productive effort is not re-

Figure 30-2. Number of stations at which ichthyoplankton samples have been collected in spring, March-May.

Figure 30-3. Number of stations at which ichthyoplankton samples have been collected in summer, June-August.

Figure 30-4. Number of stations at which ichthyoplankton samples have been collected in fall, September-November.

Figure 30-5. Number of stations at which ichthyoplankton samples have been collected in winter, December-February.

the effort into quarter-years (Figs. 30-2-30-5) shows that 35 percent of the sampling was done in spring (March-May), 55 percent in summer (June-August), only 9 percent in fall (September-November), and less than 1 percent in winter (December-February). Thus, fish whose eggs or larvae are planktonic during the six months from September through February were undersampled in comparison to those that are planktonic during spring and summer. Pleuronectidae, for example, show considerable diversity in time of spawning. Yellowfin sole spawn in midsummer and thus are not available for capture with plankton nets in spring or early summer. Halibut and turbot, on the other hand, spawn in winter and by the following spring and summer their larvae are difficult to capture with plankton nets. Pollock, one of the most abun-

dant fish in the Bering Sea, spawn mainly during spring and their larvae are relatively scarce in samples collected during late summer and fall.

A further obstacle to inter-cruise comparisons is the difference in types of tow used in making the collections. Vertical net tows, which usually strain a relatively small volume of water, capture only small numbers of fish larvae, and the more uncommon taxa are seldom found in collections made by this method. A horizontal tow at the surface—a neuston tow for example—may produce large numbers of eggs and larvae, but these may belong to only a few taxa whose larvae characteristically inhabit this zone. Greenlings and Atka mackerel, family Hexagrammidae, and sculpins of the genus *Hemilepidotus* are often abundant in neuston collections, whereas

TABLE 30-2

Ichthyoplankton cruises in the Bering Sea east of 174°E, 1955-1979

Cruise ¹	Year	Vessel	Date	Lat. N. From To	Long. From To	Number Stations ²	Type of tow ³	Net	Reference
1	1955	Oshoro Maru	8-20 July	52° - 58°	175°E - 163°W	17	S	FLN	Faculty of Fisheries, Hokkaido Univ. 1957a
2	1956	Oshoro Maru	23 Jul-13 Aug	52 - 63	174 E - 168 W	34	S	FLN	Faculty of Fisheries, Hokkaido Univ. 1957b
3	1957	Oshoro Maru	7-11 July	52 - 54	174 E - 172 W	9	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1958
4	1957	Brown Bear	3-29 Aug	52 - 56	177 E - 166 W	20	O	IKMT	Aron, 1960
5	1958	Oshoro Maru	8-27 June	52 - 62	175 E - 175 W	21	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1959
6	1958	Zhemchug	29 Jun-5 Aug	53 - 64	174 E - 160 W	124	S,V	IKS-80	Musienko, 1963
7	1958	Zhemchug	19 Aug-18 Sep	55 - 63	174 E - 162 W	95	S,V	IKS-80	Musienko, 1963
8	1959	Alazeya	13-19 March	54 - 59	174 E - 165 W	60	S,V	IKS-80	Musienko, 1963
9	1959	Oshoro Maru	21 Jun-12 Jul	52 - 58	174 W - 167 W	17	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1960
10	1960	Oshoro Maru	19 Jun-13 Jul	52 - 60	175 E - 163 W	13	S	FLN, PC	Faculty of Fisheries, Hokkaido Univ., 1961
11	1961	Oshoro Maru	1-14 July	53 - 60	174 E - 172 W	11	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1962
12	1962	Ogon	21-31 July	58 - 63	174 W - 163 W	88	S,V	IKS-80 F/r-80/113	Kashkina, 1965, 1970
13	1963	(Halibut Comm. charter)	May	54 - 55	166 W - 165 W	10	--	--	Dunlop et al, 1964
14	1963	Oshoro Maru	14 Jun-14 Jul	54 - 59	177 E - 167 W	7	S	FLN, IKMT	Faculty of Fisheries, Hokkaido Univ., 1964
15	1965	Seskar	March-May	54 - 61	179 W - 160 W	95	S,V,T	IKS-80	Serobaba, 1968 (report only pollock)
16	1965	Seskar	June-July	54 - 65	180 - 163 W	86	S,V,T	IKS-80	Serobaba, 1968 (report only pollock)
17	1966	Oshoro Maru	13 Jun-31 Jul	53 - 65	177 E - 164 W	32	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1967
18	1967	Oshoro Maru	12 Jun-20 Aug	53 - 65	175 E - 161 W	22	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1968
19	1968	Oshoro Maru	12 Jun-8 Aug	52 - 64	177 E - 162 W	28	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1969
20	1969	Oshoro Maru	10 Jun-2 Aug	53 - 58	180 - 160 W	35	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1970
21	1970	Oshoro Maru ⁴	6 Jun-16 Aug	52 - 58	172 W - 161 W	50	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1972
22	1971	George B. Kelez	26 May-9 Jun	53 - 56	173 W - 168 W	6	O	B-60	Dunn and Naplin 1973
23	1971	Oshoro Maru ⁴	17 Jun-23 Jul	52 - 58	174 E - 161 W	31	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1973
24	1972	Oshoro Maru ⁴	11 Jun-12 Aug	52 - 65	175 W - 161 W	--	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1974
25	1972	Professor Deryugin	March-Jun	54 - 62	174 E - 164 W	--	O,H	IKMT (CL)	Serobaba 1974 (pollock-vertical sections)
26	1973	Oshoro Maru ⁴	11 Jun-23 Jul	52 - 59	180 - 160 W	36	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1975
27	1974	Oshoro Maru ⁴	7 Jun-22 Jul	52 - 58	178 W - 167 W	41	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1976a
28	1975	Shunyo Maru	4 Mar-3 Jul	54 - 61	161 W - 179 W	121	V	1.4mLN	Wakabayashi, Mito and Nagai, 1977
29	1975	Discoverer	20 May-15 Jun	53 - 60	158 - 175 W	108	V,O	1m,NIO	NODC Data ⁵
30	1975	Oshoro Maru ⁴	13 Jun-5 Jul	52 - 59	180 - 160 W	78	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1976b
31	1975	Discoverer	15-26 Aug	54 - 60	159 - 172 W	76	V,O	1m,NIO	NODC Data ⁵
32	1975	Miller Freeman	13-24 Nov	54 - 62	162 - 170 W	38	V,O	1m,NIO	NODC Data ⁵
33	1976	Surveyor	17 Mar-26 Apr	54 - 58	162 - 174 W	57	V,O	1m,NIO	NODC Data ⁵
34	1976	Miller Freeman	26 Apr-31 May	54 - 59	175 W - 158 W	56	S,O	Sam., B-60	Waldron and Favorite, 1977
35	1976	Oshoro Maru ⁴	14 Jun-25 Jul	52 - 64	174 E - 162 W	39	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1977
36	1976	Discoverer	5-17 Aug	55 - 66	161 - 168 W	80	V,O	1m,NIO	NODC Data ⁵
37	1977	Miller Freeman	16 Apr-13 May	53 - 58	174 W - 163 W	134	S,O	Sam., B-60	Waldron and Vinter, 1978
38	1977	Oshoro Maru ⁴	12 Jun-24 Jul	52 - 60	178 W - 163 W	51	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1978
39	1978	Miller Freeman	11 Feb-18 Mar	54 - 60	164 W - 177 W	32	S,O	Sam., B-60	Waldron, 1978
40	1978	T. G. Thompson	11-29 Apr	--	--	157	S,O	N, B, NIO	PROBES, 1979 ⁶
41	1978	Ogon	10 Apr-20 May	54 - 62	162 W - 180	349	S,V	X-80	Moiseev and Bulatov, 1979 ⁷
42	1978	Oshoro Maru	June-August	--	--	66	S	FLN	Faculty of Fisheries, Hokkaido Univ., 1979
43	1979	Hakuho Maru	23-30 July	55 - 65	168 W - 173 W	5	S,O	ORI-100	Haryu, Endo and Nishiyama, 1979

¹ Cruise No.—an arbitrary number assigned for the purposes of this summary.² Type of tow: S=horizontal surface tow probably to a maximum depth of 1 m; O=oblique tow from the surface to a depth and back to the surface; V=vertical net tow; H=a horizontal net tow at some depth below 1 m; T=Total; definition of this not known, but the terminology appeared in Serobaba (1968).³ Net: FLN—fish larva net. See references for specifications which varied slightly between 1955 and 1977.

IKMT—Isaacs-Kidd Midwater Trawl (Isaacs and Kidd 1953).

IKMT(CL)—an IKMT with a closing type cod end.

IKS-80—"...nets of 140 mesh with a mouth diameter of 80 cm." (Kashkina 1970).

B/r-80/113—a reverse cone net with dimensions and mesh the same as the IKS-80.

B—bongo net, size not specified.

B-60—bongo net with a mouth diameter of 60 cm and mesh of 0.505 mm.

1m—a simple one-meter ring net.

N—a neuston net.

Sam.—A Sameoto type sampler (Sameoto and Jaroszynski 1969).

NIO—a 2-meter National Institute of Oceanography trawl.

PC—underway plankton catcher.

⁴ Reports of the *Oshoro Maru* cruises since 1969 do not list station positions or catch, but give only total stations sampled.⁵ Catch by station for these cruises was obtained from the OCSEAP data file at NODC.⁶ Cruise report PROBES 78, Leg I, 8 April to 1 May 1978.⁷ Moiseev, E. I. and O. A. Bulatov 1979. The state of pollock stocks in eastern Bering Sea. Pacific Research Institute of Fisheries and Oceanography (TINRO), Vladivostok 1979.

flatfish larvae are almost completely absent from such tows. In general, an oblique tow from near bottom to the surface appears to capture a large number of species as well as large numbers of specimens. Examination of Table 30-1 shows that many of the collections, including most of those done aboard the *Oshoro Maru*, were made with surface tows. Because those cruises represent a majority of sampling efforts west of the continental slope, the species composition in that area is biased towards those taxa with surface-dwelling planktonic larvae.

A third point to consider in comparing samples collected during different cruises is the area coverage and station spacing. Economics, logistics, and program objectives usually dictate that a particular sampling effort be confined to a single vessel often operating within a small area and/or sampling at relatively few stations. The *Oshoro Maru* cruises often covered a large area but with widely spaced stations. Most of the effort by the *Miller Freeman* consisted of sampling at closely spaced stations within a relatively small area. Cruises by the *Zhemchug* and *Seskar* covered large areas within which stations were spaced at moderate intervals, but most sampling was restricted to the continental shelf and slope.

As a result of restricted areal coverage combined with uneven seasonal distribution of effort and the limitations of the type of tow that predominated in certain areas, the distribution of many species is not well defined. Yellowfin sole larvae are not reported in very many collections; this may be because their spawning area appears to be well up on the continental shelf in moderately shallow water, whereas much of the sampling effort was devoted to the outer continental shelf and continental slope. Pacific herring spawn in intertidal areas and their larvae must remain close to shore while they are planktonic, for they are reported in very few collections. Pollock spawn along the continental slope and outer shelf and their eggs and larvae are seldom captured in water of less than 75 m depth. Myctophids and bathylagids are probably undersampled because of the predominance of surface tows in the central basin where they inhabit the deeper waters.

Much of the sampling effort, then, has been focused on the distribution and abundance of a few species, in particular pollock and yellowfin sole; collections of other species have been incidental to this main objective. It is likely that definitive information about the ichthyoplankton community in the Bering Sea can be gained only through a concerted cooperative effort by Japanese, Soviet, and U.S. biologists to promote at least one year-round sampling effort covering the entire Bering Sea, with

sampling at moderately spaced stations, and including at least one oblique and one neuston tow at each station.

AVAILABLE FAUNA AND CATCH

There are generally considered to be about 300 species of fish in the Bering Sea, divided among 150 genera and 40-45 families (Quast and Hall 1972, Wilimovsky 1974). Of these, four families—Petromyzontidae, Squalidae, Rajidae, and Acipenseridae—with 6 genera and 12 species, are extremely unlikely to appear in plankton samples. In addition, larvae of Salmonidae and Gasterosteidae, with 7 genera and 16 species, are seldom found in marine plankton, although they do occur in samples from estuaries and brackish water areas. There remain 34 families with 137 genera and 270 species that can reasonably be expected to occur in plankton samples either as eggs or larvae, or both.

Collections resulting from the cruises discussed here contain 87 separate taxa, of which 60 are species, 13 are genera (9 of these are also included in the specific listing), and 14 are families. All told, they include representatives of 52 genera in 24 families; the various taxa and the collections from which they were obtained are presented in Table 30-2. Thus, larvae of over one-third of the genera, over half of the families, but of only about one-fifth of the species of fish present in the Bering sea as adults have been collected and identified. Most of the species not included in the list are in the large families Cottidae, Zoarcidae, Cyclopteridae, and Stichaeidae, which account for some 170 species. The 14 families not included in the collections contain only 23 species.

Because of the variations in spawning time and the logistics of vessel cruises, it is seldom possible to sample the entire spectrum of ichthyoplankton during any single cruise. Examination of Table 30-2 shows that of a total of 87 taxonomic categories of larvae, no report lists more than 38 taxa and four reports list 10 or fewer taxa. Certain of the reports in Table 30-1 (e.g., Serobaba 1968, reporting on the 1965 cruise of the *Seskar*) deal with single species and hence do not include listings of other larvae and eggs that were almost certainly present in the collections.

DISTRIBUTION AND ABUNDANCE OF ICHTHYOPLANKTON

*Clupeidae*¹

The Pacific herring, *Clupea harengus pallasi*, is the only member of this family in the Bering Sea; as an

¹The arrangement of families presented here follows that proposed by Greenwood et al. (1966).

adult it is abundant and commercially valuable, but larvae were caught at only 10 stations on three cruises in the summer and fall of 1975 and the summer of 1976. Only 24 herring larvae were caught, all at shallow water stations in Bristol Bay and Norton Sound (Fig. 30-6), and the largest catch was 9 larvae. A partial explanation of the scarcity of herring larvae is that spawning takes place in intertidal areas and larvae apparently remain close to shore during the period when they would be susceptible to capture with plankton nets. Most of the sampling effort shown in Fig. 30-1 was in areas outside those occupied by herring larvae.

Herring eggs are demersal and none were reported.

Osmeridae

Four species of smelt occur in the Bering Sea as adults and two of these were present in plankton samples. Capelin, *Mallotus villosus*, were caught at 76 stations on 10 cruises in spring, summer and fall (Fig. 30-7). With one exception, all occurred south of 60°N almost exclusively over the continental shelf and extending into the easternmost part of Bristol Bay. The single exception was a capelin larva collected just south of Bering Strait in August 1976. Collections made aboard the *Miller Freeman* in the spring of 1977 (Waldron and Vinter 1978) indicate that capelin are more easily caught with neuston nets than with deeper fishing bongo nets; more specimens were collected at night than during the day.

A single juvenile Rainbow smelt, *Osmerus mordax*, was caught in quadrangle 61½°N, 167°W in November 1975.

Unidentified osmerid smelt were reported at five stations on two cruises in spring and summer.

Osmerid smelt are demersal spawners and no eggs were reported.

Bathylagidae

The deepsea smelts are represented in the Bering Sea by five species (one name is of questionable validity), and larvae of three of the species were found in plankton samples. The most frequently collected species was the northern smoothtongue, *Bathylagus schmidtii*, which was caught at 69 stations on 10 cruises in spring, summer, fall, and winter. Catches of *B. schmidtii* were generally fewer than 5 larvae per tow with a maximum of 22 at one station.

The Pacific blacksmelt, *Bathylagus pacificus*, was collected at 51 stations on six cruises in spring and summer. Catches of *B. pacificus* were usually fewer than five larvae per tow with a maximum of eight at one station. The stout blacksmelt, *Bathylagus milleri*, was collected at only two stations on one cruise in summer; the two specimens should be described as large juveniles or small adults even though they were collected with a plankton trawl. In contrast to Osmeridae, deepsea smelts were collected over the outer continental shelf, the continental slope, and the deep central basin south of 60°N and from 165°W to 178°E (Fig. 30-8). Since there did not appear to be any marked difference in distribution between the northern smoothtongue and the Pacific blacksmelt, their distribution was plotted as one figure. The two stout blacksmelt were collected at two stations in quadrangle 52½°N, 178°E. As far as can be determined, none of the bathylagid larvae were captured with surface nets. Because most of the sampling effort over the deep central basin was with

Figure 30-6. Number of stations at which larvae of various taxa have been caught. Each symbol represents one station.

Figure 30-7. Number of stations at which larvae of Osmeridae have been caught.

surface tows, the lack of specimens from that portion of the Bering Sea may be a sampling artifact.

Although eggs of deepsea smelt are pelagic and identifiable, none were reported.

Chauliodontidae

Of this family of bathypelagic fish, a single species exists in the Bering Sea, the Pacific viperfish, *Chauliodus macouni*. Larvae and juveniles were collected at only three stations on two cruises in summer, all of which were close to the Aleutian Islands between 172°W and 178°E (see Fig. 30-6).

Eggs of this species are pelagic but none were reported.

Myctophidae

The family of lanternfishes includes seven species in the Bering Sea; larvae of three of these were reported from plankton collections. Larvae of the northern lampfish, *Stenobrachius leucopsarus*, were collected most frequently of the three species and were caught at 35 stations on nine cruises in spring, summer, and fall. Northern lampfish larvae were distributed from the edge of the continental shelf across the slope and along the Aleutian Islands westward to 177°E (Fig. 30-9). Only one larva of this species was caught north of 57°N, and there were no large catches made at any station.

Larvae of the bigeye lanternfish, *Protomyctophum thompsoni*, were collected at 10 stations on three cruises in spring and summer. They were caught at two stations along the continental shelf and at eight stations close to the Aleutian Islands westward to 177°E. Catches were small, consisting of only one or two larvae at any one station.

Figure 30-8. Number of stations at which larvae of Bathylagidae have been caught.

California headlightfish, *Diaphus theta*, were caught at only four stations on one cruise in summer. These specimens were juveniles or young adults found close to the Aleutian Islands from 174°W to 178°E.

Almost all myctophid larvae were caught with obliquely or vertically towed nets that fished well below the surface. As with bathylagids, the apparent absence of myctophid larvae from the central deep-water region of the Bering Sea may be an artifact caused by surface sampling rather than an actual distribution pattern.

Little is known about eggs of myctophids and although they are assumed to be pelagic, none were reported.

Gadidae

There are four or possibly five species of codfish in the Bering Sea, and larvae of three of these were found in 21 plankton collections made during all seasons.

Larvae and juveniles of saffron cod, *Eleginus gracilis*, were caught at 13 stations on two cruises in August 1975 and 1976. All of those classified as larvae were caught in Norton Sound and Bering Strait; the juveniles were caught in outer Bristol Bay (Fig. 30-10). Most samples contained only one or two specimens, but one sample from quadrangle 63½°N, 164°W yielded 54 saffron cod larvae.

Eggs of saffron cod are demersal and none were reported.

Larvae of Pacific cod, *Gadus macrocephalus*, were caught at five stations on three cruises in spring and summer. The total catch was only 10 larvae, five from a station in quadrangle 52°N, 174°W in the Aleutian Islands and five from four stations on the

Figure 30-9. Number of stations at which larvae of Myctophidae have been caught.

continental shelf south of Nunivak Island. They were caught in surface, oblique, and vertical net tows.

Eggs of Pacific cod are demersal or benthonic and none were reported.

Larvae identified only to the family Gadidae were caught at 41 stations on eight cruises in spring and summer. Catches at individual stations usually amounted to fewer than five larvae, but at one station just west of Unimak Pass, 59 larvae of this taxon were caught during a spring cruise. The distribution of larvae identified only as Gadidae extended from the Aleutian Islands to Bering Strait and from 163°W to 176°W. Stations at which these larvae were caught were scattered from well up on the continental shelf across the slope and, rarely, over deep water. It seems likely that the unidentified gadids from north of St. Lawrence Island belonged to a different taxon from those to the south.

Walleye pollock, *Theragra chalcogramma*, have probably been the primary objective of more ichthyoplankton surveys than any other fish in the Bering Sea. As a result the distribution and abundance of their eggs and larvae are quite well known and have been described in various reports (Musienko 1963; Serobaba 1968, 1974; Waldron and Favorite 1977; Waldron and Vinter 1978; Waldron 1978; Nishiyama, Volume 2 of this book). These reports show larvae to be most abundant between Unimak Pass and the Pribilof Islands along the continental slope, with total distribution extending to 59°N in spring. During summer, larvae show a more widespread distribution from the Aleutian Islands to 64½°N, and from well up on the continental shelf in Bristol Bay across the central basin to 177°E.

Because of the large number of samples containing

larvae it was possible to show distribution for two periods, spring and summer, and for two types of net tows, surface tows and combined oblique and vertical tows. Larvae were caught in surface tows made at 40 stations on two cruises in March-May, and at an undetermined number of stations on one other spring cruise for which the catch by station could not be determined. These surface samples showed the larvae to be distributed along the continental slope in a somewhat discontinuous pattern (Fig. 30-11). During the same two cruises larvae were caught in oblique net tows at 124 stations and on another spring cruise at 15 stations. Distribution shown by these oblique tows covers essentially the same areas as shown by the surface tows but there was more continuity (Fig. 30-12).

During June-August, larvae were caught in surface tows at 76 stations, mainly on nine cruises of the *Oshoro Maru*, and were shown to be widely distributed over the entire Bering Sea (Fig. 30-13). Fewer vertical and oblique tows were made during summer and larvae were caught at only 31 stations; yet these showed a wide distribution from the Aleutian Islands to 63½°N and from 164°W to 178°E (Fig. 30-14).

Sampling for larvae has been intensive along the continental slope and outer shelf, but there is a gap in our knowledge of their distribution and abundance over the deep central basin. Comparison of catches in surface and oblique net tows shows that relatively fewer larvae are caught in surface tows and at fewer stations. Most of the sampling over the central basin was with the surface tows (see Figs. 30-1 and 30-3) and most of those were during summer. It is reasonable to assume that sampling by means of oblique tows would reveal a larger population of pollock

Figure 30-10. Number of stations at which larvae of Gadidae have been caught.

Figure 30-11. Number of stations at which larvae of *Theragra chalcogramma* have been caught with surface tows in spring, March-May.

Figure 30-12. Number of stations at which larvae of *Theragra chalcogramma* have been caught with oblique or vertical tows in spring, March-May.

Figure 30-13. Number of stations at which larvae of *Theragra chalcogramma* have been caught with surface tows in summer, June-August.

Figure 30-14. Number of stations at which larvae of *Theragra chalcogramma* have been caught with oblique and vertical tows in summer, June-August.

Figure 30-15. Area within which eggs of *Theragra chalcogramma* have been caught, and centers of abundance in five years.

larvae in the deep central basin than is shown by surface tows.

Since pollock eggs are among the few routinely identified in plankton samples, their distribution can be shown in some detail. They are present generally along the outer continental shelf and continental slope from the Aleutian Islands to 60°N (Fig. 30-15). Within this area they occur from late February to late July but are most abundant from about mid-March to mid-May. Abundance of eggs has been treated in greater detail than that of larvae, probably because eggs are caught more readily than larvae in the vertical net tows which were used on most of the Soviet cruises. Quantitative estimates of

abundance are shown for cruises made in 1959 (Musienko 1963), in 1965 (Serobaba 1968), and in 1972 (Serobaba 1974), although the last shows only vertical distribution along section lines. Estimates of egg abundance based on oblique net tows are given for 1976 (Waldron and Favorite 1977), for 1977 (Waldron and Vinter 1978), and for 1978 (Waldron 1978). Maximum abundance of eggs beneath 1 m² of sea surface, as shown by vertical net tows, was 598 in March 1959 between Unimak Pass and the Pribilof Islands and over 2,000 eggs at some time from March to May 1965 immediately north of Unimak Pass. Maximum numbers of eggs beneath 1 m² shown by oblique net tows was 1,268 in May

1976 east of the Pribilof Islands, 1,041 in April 1977 north of Unimak Island, and 958 in March 1978 northwest of Unimak Pass.

Since most of the samples of ichthyoplankton were collected with open nets that caught all plankton between the maximum tow depth and the surface, there is little information on the vertical distribution of the various fish larvae and eggs. During a Soviet cruise in 1972 (Serobaba 1974), samples were collected with a closing net fished at several different levels. Results of this survey show pollock eggs and larvae present to depths of 1,000 m, but catches below 100 m were small, and below 200 m they were fewer than one per 1,000 m³.

Zoarcidae

Eelpouts are one of the larger families with 29 species present as adults in the Bering Sea, yet larvae have been recorded for only three stations (see Fig. 30-6). All eelpouts were caught at stations over the continental shelf during spring, summer, and winter.

The lack of eelpouts in plankton collections probably reflects spawning characteristics, behavior of young, and configuration of the net tow used to collect most plankton samples. According to Hart (1973), reproduction among eelpouts may be oviparous or ovoviviparous; Clemens and Wilby (1961) state that eelpouts may be viviparous. Eggs of some eelpouts are as large as 7 mm (*Lycodes palearis*, in Hart 1973) and hence the newly hatched young might be quite large, say on the order of 20 mm, and active. If large size and active behavior were coupled with dwelling on or close to the bottom, larvae of eelpouts would be difficult to capture with standard plankton nets. In order to determine the distribution of larval eelpouts it is likely that a special sampler will have to be used to fish the layer immediately above bottom.

Macrouridae

Rattails or grenadiers are represented in the Bering Sea by nine species, and *Coryphaenoides pectoralis* is thought to be abundant enough to support a commercial fishery (Novikov 1970). Larvae of grenadiers were caught at eight stations on five cruises in spring and summer, all close to or over the continental slope between the Aleutian Islands and 56½°N, and between 167°W and 180° (see Fig. 30-6); only one or two larvae were caught at any station.

Macrourid eggs are pelagic and identifiable to family, and were caught in a vertical tow near 55°N, 177°E (Musienko 1963).

Melamphaeidae

This family of bathypelagic fish is represented in the Bering Sea only by the highsnout melamphid, *Melamphaes lugubris*; two small adults were caught at one station in quadrangle 52½°N, 178°E in the western Aleutian Islands (see Fig. 30-6).

No melamphid eggs were reported.

Scorpaenidae

Rockfish are a commercially important group with 12 species of *Sebastes* and 3 of *Sebastolobus* in the Bering Sea. At present it is difficult, if not impossible, to identify larval rockfish from the Bering Sea to species although it is possible to separate genera. Larval *Sebastolobus* were not reported from any of the plankton collections, but *Sebastes* sp. larvae were caught at 89 stations on 10 cruises made in spring and summer. Individual catches were not large, generally fewer than 20 larvae per station, but a few samples contained 30-40 larvae each. Rockfish larvae were distributed mainly over the continental slope, the adjacent shelf, and deep water, although a few were caught well up on the continental shelf. They were caught from the Aleutian Islands north to about 63°N, although only two stations were north of 60°N, and from 164°W westward to 175°E (Fig. 30-16). Rockfish larvae were collected with surface, oblique, and vertical net tows, but most appear to have been caught in oblique tows.

Rockfish of the genus *Sebastes* are ovoviviparous and no eggs were reported.

Hexagrammidae

The greenlings and Atka mackerel were one of the most commonly occurring groups of larvae in all plankton samples. In the Bering Sea the family contains two genera—*Pleurogrammus*, with one species, the Atka mackerel, *P. monopterygius*, and *Hexagrammos* or greenlings, with five species. Since identification to species of the greenlings is not always possible, these larvae have been treated here only as a genus. Greenling larvae were collected at 251 stations on 21 of the 26 cruises; only one other genus occurred as frequently and that was the sculpin genus *Hemilepidotus*. Greenling larvae appeared in catches made during all seasons of the year from late February to November. Catches at individual stations were moderate, usually fewer than 10 larvae, and the maximum catch was 253 larvae at a station in quadrangle 53°N, 178°E in June 1960. Larval *Hexagrammos* sp. were distributed widely across the Bering Sea from the Aleutian Islands to 62°N and from 161°W in Bristol Bay to 174°E, but appeared to

be caught more frequently south of the Pribilof Islands along the edge of the continental shelf and to the west in the vicinity of Bowers Ridge (Fig. 30-17).

Larvae of Atka mackerel were caught at 118 stations on 11 cruises in winter, spring, and summer from late February to August. Catches at individual stations were generally smaller than for greenling larvae and the maximum catch was 37 larvae at a station in quadrangle 55°N, 170°W. Atka mackerel larvae were distributed over much the same area as were greenlings, occurring from the Aleutian Islands to 62°N and from 160°W in Bristol Bay to 175°E, but they were present at less than half as many stations as greenlings (Fig. 30-18). Both greenlings and Atka mackerel larvae are caught more frequently with surface than with oblique and vertical net tows.

Eggs of Hexagrammidae are demersal and none were reported.

Figure 30-16. Number of stations at which larvae of *Sebastes* sp. have been caught.

Figure 30-18. Number of stations at which larvae of *Pleurogrammus monopterygius* have been caught.

Anoplopomatidae

The family of sablefishes is composed of two species, both of which are present in the Bering Sea as adults.

One larva tentatively identified as skilfish, *Erilepis zonifer*, was caught at 57°41'N, 174°25'W in July 1960.

Larvae of sablefish, *Anoplopoma fimbria*, were caught at 28 stations on 11 cruises in spring and summer. Usually only a few were caught at any one station, but an unusually large catch of 45 larvae was made in quadrangle 55°N, 164°W near Unimak Island in 1968. Larvae of sablefish were distributed generally from the Aleutian Islands to about 59°N and from 162°W to 175°E, but were found most frequently along the edge of the continental shelf and in the vicinity of Bowers Ridge in the central Aleutian Islands (Fig. 30-19). More appear to have been

Figure 30-17. Number of stations at which larvae of *Hexagrammos* sp. have been caught.

Figure 30-19. Number of stations at which larvae of *Anoplopoma fimbria* have been caught.

caught with surface than with oblique or vertical net tows.

Eggs of sablefishes are reportedly pelagic, but none were found.

Cottidae

With some 75 species in 36 genera, sculpins are by far the largest family of fish in the Bering Sea and were present in all 26 collections. Larvae of this group can usually be identified as cottids but relatively few are identifiable to species. In these collections only seven genera with six species were listed, the remainder coming under the heading of unidentified cottids (see Table 30-2). Of these, three species and one genus each were captured on only one cruise.

The most commonly occurring and widespread sculpin larvae were the Irish lords, *Hemilepidotus* sp., which were caught at 208 stations on 21 cruises during all seasons of the year from late February to November. Irish lords include five species in the Bering Sea, one of which is found only on the western side. Identification to species is possible for larvae with countable finrays but the smaller larvae are usually identified only to genus; thus distribution is shown only for the genus. Irish lords have one of the widest distributions of all larvae in the Bering Sea, extending from the Aleutian Islands to 63°N and from 160°W in Bristol Bay to 174°E (Fig. 30-20). Although there are insufficient quantitative data for the entire region, collections made aboard the *Oshoro Maru* during summer were all made by approximately the same type of surface tow, and these show no areas of consistently high abundance. Actual catches of Irish lord larvae for a 10-minute surface tow range from 1 to 3,709, with only 5 of 114 collections

containing more than 100 larvae. The catch of 3,709 larvae was made in quadrangle 58½°N, 165°W. Only the northeastern area of the Bering Sea, roughly the Nunivak Island/St. Matthew Island/St. Lawrence Island area, appears to be devoid of Irish lord larvae. This genus of sculpins appears to share the surface layer with the hexagrammid larvae, for more are caught with neuston and other surface-towed nets than with nets sampling deeper waters. Irish lords differ from the hexagrammids in that there is a greater difference in day and night catches of the sculpins than of hexagrammids, with larger catches made at night.

A second genus of sculpins common in these collections was *Myoxocephalus*, great sculpins, with 10 species present in the Bering Sea. Since the taxonomy of the adults is unclear, no attempt was made in any of the collections to identify larvae to species. Larvae of great sculpins did not occur as frequently as those of Irish lords but were caught at 53 stations on nine cruises in the summer. They have one of the widest north-south distributions, extending from the Aleutian Islands to 65½°N, but appear to be limited mainly to the continental shelf, with east-west distribution extending from 161°W to 174°E (Fig. 30-21). Although the catches were small, one or two larvae per station, the great sculpins are present in the northeastern area in which Irish lords were absent. The area of greatest abundance is in Bristol Bay, where catches of up to 408 larvae per station were reported in quadrangle 57°N, 161°W, and catches of more than 50 larvae were reported by several stations. Most of the great sculpin larvae were caught with surface tows.

Figure 30-20. Number of stations at which larvae of *Hemilepidotus* sp. have been caught.

Figure 30-21. Number of stations at which larvae of *Myoxocephalus* sp. have been caught.

In addition to *Hemilepidotus* sp. and *Myoxocephalus* sp., six other taxa of sculpins, plus a category called unidentified cottids, occurred in the samples. Distribution of these miscellaneous species is shown in Fig. 30-22.

Artediellus pacificus, the hookhorned sculpin, was collected at one station east of the Pribilof Islands on the continental shelf during late spring.

Icelinus borealis, the northern sculpin, was caught at two stations between Unimak Island and the Pribilof Islands on one cruise in spring.

Gymnocanthus sp. was caught at one station northeast of St. Lawrence Island during summer.

Malacocottus sp., probably *M. kincaidi*, the black-fin sculpin, was caught at four stations between Unimak Pass and the Pribilof Islands during a spring cruise.

Triglops pingeli, the ribbed sculpin, was caught at one station southeast of Nunivak Island during a fall cruise, and *Triglops* sp. was caught at three additional stations, one northeast of St. Lawrence Island on a summer cruise, one between Unimak Pass and the Pribilof Islands, and one near the Alaska Peninsula during a spring cruise.

In addition to these identifiable taxa, there were numerous cottids which could only be identified to family. These were distributed mostly over the continental shelf and slope south of 60°N, but a few were collected as far north as Bering Strait and west along the Aleutian Islands to 177°E. Unidentified cottids were collected at 83 stations on 11 cruises in spring and summer.

Eggs of cottids are demersal and none were found in any of the plankton collections.

Figure 30-22. Number of stations at which larvae of various cottids have been caught. Each unnumbered symbol represents one station.

Agonidae

The poachers are represented in the Bering Sea by 16 species and larvae of seven of these were present in the plankton samples. Larval poachers were caught at 45 stations on 14 cruises in spring and summer (Fig. 30-23).

Agonus acipenserinus, the sturgeon poacher, was caught at 18 stations on seven cruises, all of which were in summer. They were distributed mainly in Bristol Bay and the adjacent continental shelf except for one larva caught at a station north of Nunivak Island.

Odontopyxis trispinosa, the pygmy poacher, was caught at four stations on one cruise in summer, all north of St. Lawrence Island and in Bering Strait.

Asterotheca infraspinata, the spinycheek starsnout, was caught at two stations just north of Unimak Island in summer.

Aspidophoroides sp. larvae were caught on two cruises, *A. bartoni* on one cruise, and *A. olriki* on another, all in summer. Larvae of this genus were found north of St. Lawrence Island (*A. olriki*) and along the continental shelf edge south of the Pribilof Islands.

Hypsagonus quadricornis, the fourhorn poacher, was caught on two cruises in spring and summer, with one specimen from the Gulf of Anadyr and the other specimen north of Unimak Island.

Ocella dodecaedron, the Bering poacher, was caught at one station in summer on the edge of the continental shelf between Unimak Pass and the Pribilof Islands.

Larval agonids identified only to family were caught at 22 stations on six cruises in spring and

Figure 30-23. Number of stations at which larvae of various agonids have been caught. Each unnumbered symbol represents one station.

summer. They were distributed from the Aleutian Islands north to 58°N, and from 161°W in Bristol Bay to 174°W along the Aleutian Islands and along the continental slope.

Eggs of agonids are demersal and none were reported.

Cyclopteridae

The snailfish and lumpsuckers, with 49 species, are the second largest family (after Cottidae) in the Bering Sea and one of the least known with respect to both adult and larval identification. Cyclopterids were present in samples from 20 cruises, but only four taxa were identified, in addition to larvae identified to family level.

Aptocyclus ventricosus, the smooth lumpsucker, was caught at five stations on four cruises with a total catch of six specimens. All were caught with surface nets during spring and summer, four of them near the Aleutian Islands between 166°W and 175°W, and two specimens at a station in quadrangle 55½°N, 171°W (Fig. 30-24).

Nectoliparis pelagicus, the tadpole snailfish, was caught at 10 stations on four cruises during spring and summer, with a total catch of 20 specimens. Most were caught in nets fished below the immediate surface, but a few were taken with surface nets. The tadpole snailfish was found in a relatively small area between the Aleutian Islands and 56°N, and between 168°W and 175°W.

Larvae of snailfish of the genus *Liparis*, of which there are 13 species in the Bering Sea, were caught at 41 stations on 12 summer cruises. Most samples contained relatively few larvae but at one station in quadrangle 57½°N, 162°W, 671 larval *Liparis* sp.

were caught (see Cruise 20, Table 30-1). Larvae of this genus were distributed widely from the Aleutian Islands to Bering Strait and between 162°W and 175°E. They were caught most frequently over the continental shelf but were also found over the continental slope and adjacent deep water. Most of them were caught with surface nets.

Unidentified snailfish were caught at 46 stations on 10 cruises during spring and summer, between the Aleutian Islands and 60°N, and from 162°W to 174°E. Within these limits they were found over the same areas as were *Liparis* sp.

Cyclopterid eggs are demersal and none were reported.

Bathymasteridae

The searchers and ronquils are represented in the Bering Sea by three species in two genera, with one genus and one species present in the samples. Bathymasterids were caught at 76 stations on 18 cruises in spring, summer, and fall—only three specimens were caught in fall. The family is widespread in the Bering Sea; it occurred at stations from the Aleutian Islands to 63°N and from 163°W to 174°E (Fig. 30-25). Specimens were found over the continental shelf, slope, and deep water of the central basin; more were collected with surface nets than with nets that fished deeper. Catches at individual stations were generally small—fewer than five specimens—but at two stations (in quadrangles 53°N, 174°E and 55½°N, 167°W) 262 and 245 were caught (Cruise 18, Table 30-1). Only six specimens of *Bathymaster signatus*, the searcher, were caught, three each at two stations in quadrangles 54½°N, 169°W and 55°N, 167°W.

Figure 30-24. Number of stations at which larvae of Cyclopteridae have been caught. Each unnumbered symbol represents one station.

Figure 30-25. Number of stations at which larvae of Bathymasteridae have been caught.

Eggs of bathymasterids are demersal and none were reported.

Anarhichadidae

There are one or possibly two species of wolffishes in the Bering Sea; they were represented in the collections by five larval *Anarhichas orientalis*, the Bering wolffish, caught at five stations on one cruise in spring (see Cruise 37, Table 30-1). All were caught with neuston nets at stations near the outer edge of the continental shelf between Unimak Pass and the Pribilof Islands (see Fig. 30-6).

Eggs of wolffish are demersal and none were reported.

Stichaeidae

The pricklebacks, a fairly large family with 21 species present in the Bering Sea, were represented in these collections by six species in four genera and stichaeid larvae identified only to family. Pricklebacks, present in 23 of the collections, had one of the widest distributions of all the fish larvae, from the Aleutian Islands to 65°N, and from 158°W in Bristol Bay to 174°E (Fig. 30-26). Most stichaeid larvae were caught with surface tows, although some were taken with oblique and vertical tows. One species, *Lumpenus maculatus*, was caught more frequently in oblique tows.

Alectridium aurantiacum, the lesser prickleback, was collected at only a single station (see Cruise 1, Table 30-1) in quadrangle 54½°N, 165°W.

Chirolophis polyactcephalus, the decorated warbonnet, was caught at 27 stations on two cruises in spring in the general area between the Aleutian and Pribilof Islands from deep water to the outer conti-

mental shelf. Catches were moderate, usually fewer than 20 larvae per station, with a maximum catch of 78 at a station in quadrangle 53½°N, 168°W.

The genus *Lumpenus* included those pricklebacks identified only to genus, plus three species: *L. fabricii*, the slender eelblenny; *L. maculatus*, the daubed shanny; and *L. medius*, the stout eelblenny. *Lumpenus* sp. was caught at 43 stations on 10 cruises in spring and summer over an area almost as wide as for the family, from the Aleutian Islands to 63°N and from 161°W to 174°E at stations far up on the continental shelf, over the continental slope, and over the deep central basin. Catches at individual stations were generally small and the maximum catch was only 22 larvae.

Larvae of *Lumpenus fabricii* were caught at 12 stations on three cruises in summer over a wide area from the vicinity of Unimak Island to 65°N and from 160°W in Bristol Bay to 178°W in the Gulf of Anadyr. All catches of this species were made over the continental shelf and individual catches were only one or two larvae per station.

L. maculatus larvae were caught at 10 stations on three cruises in spring and summer. More restricted in distribution than *L. fabricii*, they were found from the Alaska Peninsula to only 58½°N and from 161°W to 173°W. All larvae of this species were caught over the continental shelf and, in contrast to most of the other stichaeids, more larvae of *L. maculatus* were caught in oblique and vertical tows than in surface tows. Individual catches usually consisted of only one or two larvae, with a maximum catch of 19 at a station in quadrangle 58½°N, 170°W.

A single larva of *L. medius* was caught in summer at a station in quadrangle 58°N, 166°W.

Larvae of *Stichaeus punctatus*, the arctic shanny, were caught at 18 stations on four cruises in spring and summer. They had a fairly wide distribution over the continental shelf, continental slope, and deep water from 54°N to 62½°N and between 164°W and 177°E. Individual catches were small and variable, ranging from a few to moderate numbers with a maximum of 131 larvae at a station in quadrangle 62½°N, 172°W.

Stichaeid larvae identified only to family were caught at 69 stations on 11 cruises in spring and summer. They were found over a range extending from the Aleutian Islands to 60°N and from eastern Bristol Bay at 158°W to 174°E, and over the continental shelf, continental slope, and deep water of the central basin. Individual catches ranged from one or two to a maximum of 556 per station, with the maximum caught at a station in quadrangle 54½°N, 170°W.

Figure 30-26. Number of stations at which larvae of Stichaeidae have been caught. Each unnumbered symbol represents one station.

Stichaeids are demersal spawners and no eggs were reported.

Cryptacanthidae

Of the two species of wrymouths in the Bering Sea, one, *Lyconectes aleutensis*, the dwarf wrymouth, was represented in these collections by a larva caught during an early spring cruise at a station in quadrangle 55°N, 165°W (see Fig. 30-6).

The wrymouths are demersal spawners and no eggs were reported.

Ptilichthyidae

This family contains a single species, *Ptilichthys goodei*, the quillfish; although it is not a common species, quillfish larvae were caught at 11 stations on five cruises in spring and summer. Most of the larvae were captured in Bristol Bay and out to the shelf edge, but a single specimen was caught in quadrangle 53°N, 175°E (see Fig. 30-6). All of the quillfish larvae were caught in surface tows.

Nothing is known concerning the eggs or spawning habits of quillfish.

Pholidae

The gunnells, with five species in the Bering Sea, are one of the less well known families in the area and all larvae of this taxon have been identified only to family. Gunnel larvae were caught at 40 stations on six cruises in spring and summer; most were caught during summer. They have a wide east-west distribution extending from 161°W in Bristol Bay to 174°E, but are limited in their north-south distribution from the Aleutian Islands to 58°N (Fig. 30-27). They were

taken at a few stations over the continental shelf but the majority were caught over the deep central basin. Almost all larvae were caught with nets towed at the surface; catches were often of moderate size, exceeding 50 larvae at several stations in the vicinity of Bowers Ridge; a maximum catch of 131 gunnel larvae occurred at a station in quadrangle 53½°N, 179°E.

Eggs of this family are demersal and none were reported.

Zaproridae

There is only one species of prowlfish, *Zaprora silenus*, in the Bering Sea. Larvae were caught at five stations on three cruises in spring and summer with a total catch of eight larvae. Most were caught near the Aleutian Islands over the continental slope, but one was caught near the Pribilof Islands (see Fig. 30-6).

It is not known if the eggs of this species are pelagic or demersal.

Ammodytidae

The family of sand lances is represented in the Bering Sea by one species, *Ammodytes hexapterus*, the Pacific sand lance. Larvae were caught at 170 stations on 19 cruises in spring and summer; a few were caught in early fall. They were distributed widely from the Aleutian Islands to 65½°N and from 159°W in eastern Bristol Bay to 174°E, with a majority of collections being made over the continental shelf (Fig. 30-28). Sand lances were one of the most numerous larvae in the collections, with many samples containing more than 100 larvae and a maximum catch of 1,337 larvae at a station in quad-

Figure 30-27. Number of stations at which larvae of Pholidae have been caught.

Figure 20-28. Number of stations at which larvae of *Ammodytes hexapterus* have been caught.

range $56\frac{1}{2}^{\circ}\text{N}$, 164°W . During the summer, larvae appear to be concentrated in an area just north of the tip of the Alaska Peninsula and Unimak Island, where several catches amounted to more than 400 sand lance larvae per tow. Most of the larvae were caught in surface tows, but during the spring oblique tows seemed to catch more larvae although the catches were still not large.

Sand lance are demersal spawners and no eggs were reported.

Pleuronectidae

The right-eyed flounders, one of the commercially important groups, are represented by 17 species in 14 genera in the Bering Sea. Larvae of this group were collected in 20 of the 26 cruises during spring, summer, and fall, most abundantly in spring. In general, fewer flounder larvae were caught in surface tows than in oblique or vertical tows.

The genus *Atheresthes* contains two species: *A. stomias*, the arrowtooth flounder, generally considered an eastern Bering Sea species, and *A. evermanni*, the Kamchatka flounder, considered a western Bering Sea species but possibly extending to the Pribilof Islands. Since larvae of these two species have not been described adequately to permit specific identification, the records of capture are combined and presented here as distribution of the genus. Larval *Atheresthes* sp. were collected at 74 stations on eight cruises in spring and summer. Most of these were caught over the outer continental shelf and the continental slope but a few were taken over deep water west of the slope and from shallower water in Bristol Bay (Fig. 30-29). The distribution extended from the Aleutian Islands to $58\frac{1}{2}^{\circ}\text{N}$ and

from 162°W to 177°W . Individual catches of this species were small and usually consisted of only a few larvae per tow.

Eggs of *Atheresthes* sp. are pelagic but none were reported.

The genus *Hippoglossoides* includes *H. elassodon*, the flathead sole, and *H. robustus*, the Bering flounder, two very similar species which occur in the Bering Sea with overlapping ranges. Differentiation of the adults to species is difficult and larval descriptions are not adequate to permit identification to species; data are presented here for the genus. Larval *Hippoglossoides* sp. were caught at 22 stations on seven cruises in summer and on one cruise in fall. As with all flounders, most larvae of this genus were caught in oblique or vertical tows, and catch per tow was small. *Hippoglossoides* sp. larvae were distributed from the vicinity of Unimak Pass north to 60°N and from 162°W to 173°W (Fig. 30-30). Most were caught over the continental shelf, only a few over the deeper water of the continental slope.

As with most other flounders, the eggs of *Hippoglossoides* sp. are pelagic and are readily identifiable to genus and possibly to species on the basis of diameter. *Hippoglossoides* sp. eggs were generally distributed along the outer continental shelf between about Unimak Island northwest to west of St. Matthew Island; they were found in samples collected between April and August (see cruises 6, 7, 34, and 37, Table 30-1).

Larvae of Pacific halibut, *Hippoglossus stenolepis*, were caught at 22 stations on six cruises in spring and early summer. They were usually found at stations over the continental slope or deeper water, but a few were caught on the edge of the continental shelf;

Figure 30-29. Number of stations at which larvae of *Atheresthes* sp. have been caught.

Figure 30-30. Number of stations at which larvae of *Hippoglossoides* sp. have been caught.

these were distributed in a narrow band extending from the vicinity of Unimak Pass to quadrangle 58°N, 174°W, northwest of the Pribilof Islands (Fig. 30-31). Individual catches usually consisted of only one larva but five were caught in a one-hour tow in quadrangle 54°N, 165°W (Dunlop et al. 1964).

Eggs of Pacific halibut are pelagic but are generally found at depths greater than 200 m; no halibut eggs were reported.

Larvae of *Lepidopsetta bilineata*, the rock sole, were caught at 41 stations on four cruises in spring and summer, but the catches on three of these cruises consisted of a single larva. The remaining larvae were caught in the area between Unimak Pass and the Pribilof Islands (see cruise 37, Table 30-1). Except for one rock sole larva reported from Norton Sound, this species was restricted to the continental shelf from the Aleutian Islands to 56½°N and between 163°W and 172°W (Fig. 30-32). Catches at individual stations were small; the largest catch was 12 larvae. All were caught in vertical or oblique tows.

The eggs of rock sole, unlike those of most other flounders, are non-planktonic and none were reported.

Larvae of *Limanda aspera*, the yellowfin sole, were caught at 29 stations on four summer cruises and one fall cruise. Catches were generally small, usually fewer than 20 larvae per station. Different from those of most other flounders, larvae of yellowfin sole were found only on the inshore shallow portion of the continental shelf, where they were distributed from 57½°N to 64°N and between 162°W and 170°W (Fig. 30-33).

Eggs of yellowfin sole are pelagic and large quantities were collected during several cruises (see cruises

6, 7, and 12, Table 30-1). In 1958, both eggs and larvae were caught, but in 1962 only eggs. Musienko (1963) reported maximum densities of 500-1,000 eggs/m² south of Nunivak Island in July 1958, and Kashkina (1965) reported maximum densities of 1,684 eggs/m² northwest of Nunivak Island and 336 eggs/m² south of Nunivak Island in July 1962. These are the only reports of quantitative distribution of yellowfin sole eggs.

A single postlarva of *Limanda proboscidea*, the longhead dab, was collected in early September in Bristol Bay (see Fig. 30-31). No eggs of this species were listed in any report.

Only three larvae of *Platichthys stellatus*, the starry flounder, were caught, one in the vicinity of the Pribilof Islands and two in Norton Sound (see Fig. 30-31); all were caught during summer. No eggs of this species were listed in any of the reports.

Larvae of *Pleuronectes quadrituberculatus*, the Alaska plaice, were caught at a single station north of Nunivak Island in July 1962 (see Fig. 30-31).

Eggs of Alaska plaice were caught at 57 stations on four cruises in spring and summer although during two of the cruises eggs were caught at only three stations. Eggs of this species were distributed from 55°N off Unimak Island to 59½°N and from 159°W in Bristol Bay to 175°W, and all were found over or very close to the continental shelf (Fig. 30-34). On one cruise (see cruise 34, Table 30-1) eggs of Alaska plaice were widely distributed, with centers of abundance near the Alaska Peninsula in quadrangle 55½°N, 162°W, just south of Nunivak Island in quadrangle 59°N, 167°W, and northwest of the Pribilof Islands in quadrangle 57½°N, 171°W. All the

Figure 30-31. Number of stations at which larvae of various Pleuronectidae have been caught. Each unnumbered symbol represents one station.

Figure 30-32. Number of stations at which larvae of *Lepidopsetta bilineata* have been caught.

large samples were caught with surface tows, but some eggs were caught with oblique tows.

Larvae of *Reinhardtius hippoglossoides*, the Greenland halibut or Greenland turbot, were collected at 109 stations on 10 cruises in spring and summer, occurring more frequently than any other flounder. Some were caught with nets towed at the surface but most were caught with oblique and vertical tows. Catch per station was usually small—fewer than 10 larvae per tow—but at four stations over the shelf edge and slope (see cruise 29, Table 30-1), between 57 and 102 larvae per tow were caught. They were distributed more widely than those of any other flounder—from the Aleutian Islands to the Bering Strait and from 161°W to 177°W (Fig. 30-35), from well up on the continental shelf in outer Bristol Bay to deep water beyond the continental slope.

Eggs of Greenland turbot are pelagic and can be identified, but none were reported.

DISCUSSION

Families not present in the collections

There are several families of fish represented in the Bering Sea as adults whose eggs and/or larvae were not found in plankton samples. The following list of 14 families, which include 23 species, are listed by Quast and Hall (1972) as present in the Bering Sea. Some of these may spawn in waters south of the Aleutian Islands, some may be present in deep waters of the central basin, some may be present at very shallow inshore locations, and others may be part of

the unidentified larvae component present in almost all plankton collections.

Synphobranchidae (1)	Oneirodidae (2)
Notacanthidae (4)	Zeidae (1)
Gonostomatidae (4)	Gasterosteidae (2)
Alepocephalidae (1)	Pentacerotidae (1)
Alepisauridae (1)	Trichodontidae (2)
Anopteridae (1)	Scytalinidae (1)
Moridae (1)	Bothidae (1)

Groupings of taxa by habitat

Vertical distribution of larvae could not be determined for any taxon except pollock, but certain larvae were present either exclusively or mainly in one type of tow or another. Table 30-3 shows taxa

Figure 30-33. Number of stations at which larvae of *Limanda aspera* have been caught.

Figure 30-34. Number of stations at which eggs of *Pleuronectes quadrituberculatus* have been caught.

Figure 30-35. Number of stations at which larvae of *Reinhardtius hippoglossoides* have been caught.

TABLE 30-3

Taxa of larvae caught either exclusively or predominantly in surface net tows; or in combined oblique and vertical net tows.

Surface Net Tows	Oblique or Vertical Net Tows
Osmeridae	Bathylagidae
<i>Hexagrammos</i> sp.	Myctophidae
<i>Pleurogrammos monopterygius</i>	<i>Eleginus gracilis</i>
<i>Anoplopoma fimbria</i>	<i>Gadus macrocephalus</i>
<i>Hemilepidotus</i> sp.	<i>Theragra chalcogramma</i>
<i>Myoxocephalus</i> sp.	Zoarcidae
<i>Aptocyclus ventricosus</i>	<i>Nectoliparis pelagicus</i>
<i>Liparis</i> sp.	<i>Lumpenus maculatus</i>
Bathymasteridae	<i>Lyconectes aleutensis</i>
<i>Anarchichas orientalis</i>	Pleuronectidae
Stichaeidae	
(except <i>Lumpenus maculatus</i>)	
<i>Ptilichthys goodei</i>	
Pholidae	
<i>Ammodytes hexapterus</i>	

present in two types of collections, surface tows and combined oblique and vertical tows.

It was more difficult to group larvae by water depth at stations where they were caught. In Table 30-4, six different areas are identified in which various taxa were found either exclusively or predominantly.

RESEARCH RECOMMENDATIONS

Previous cruises had some serious shortcomings: either each cruise or series of cruises focused on a particular species or area, or else the ichthyoplankton study was of secondary importance to some other primary objective. Before continuing with additional surveys of this type, it is important to obtain a comprehensive view of the total ichthyoplankton community of the Bering Sea. Lack of such information in the past has resulted in surveys with limited objectives, carried out at the wrong time of the year or in the wrong areas. The knowledge gained from a comprehensive survey would not only clarify interpretations of past fragmentary data but would also allow more economical and efficient planning of future surveys. Such a survey would almost certainly have to be a cooperative effort involving Japanese, Soviet, and United States vessels and biologists, conducted over a period of at least one year.

TABLE 30-4

Taxa of larvae caught either exclusively or predominantly in various habitats

Habitat	Taxa
Inshore	<i>Clupea harengus pallasii</i>
Inner Continental Shelf	<i>Limanda aspera</i>
Total Continental Shelf	Osmeridae <i>Eleginus gracilis</i> <i>Myoxocephalus</i> sp. Cottidae, unidentified Agonidae <i>Lumpenus fabricii</i> <i>L. maculatus</i> <i>L. medius</i> Zoarcidae <i>Ptilichthys goodei</i> <i>Lyconectes aleutensis</i> <i>Lepidopsetta bilineata</i> <i>Pleuronectes quadrituberculatus</i> <i>Platichthys stellatus</i>
Continental Shelf and Slope	<i>Theragra chalcogramma</i> (in spring) <i>Hippoglossoides elassodon</i> <i>Hippoglossus stenolepis</i>
Continental Slope and Deep Water	Bathylagidae Myctophidae <i>Anoplopoma fimbria</i> <i>Aptocyclus ventricosus</i> Pholidae <i>Atheresthes</i> sp.
Continental Shelf, Slope, and Deep Water	Gadidae, unidentified <i>Theragra chalcogramma</i> (summer) <i>Sebastes</i> sp. <i>Hexagrammos</i> sp. <i>Pleurogrammus monopterygius</i> <i>Hemilepidotus</i> sp. Bathymasteridae Cyclopteridae Stichaeidae (except those above) <i>Ammodytes hexapterus</i> <i>Reinhardtius hippoglossoides</i>

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ABSTRACT

A small but important halibut fishery exists in the Bering Sea. The distribution of halibut within the area is seasonal and dependent upon climatic conditions. The fish migrate to deep water for spawning during the winter and return to shallow areas for summer feeding. The timing and extent of the summer movement are controlled by oceanographic conditions. Spawning is known to occur between Unimak Island and the Pribilof Islands at depths of 250-550 m and probably occurs at other locations within the Bering Sea. All stages of life history forms from larvae to spawning adults have been taken.

Tagging studies have shown a movement of halibut from the Bering Sea into the Gulf of Alaska. A mixing of stocks of halibut within the Bering Sea is also suggested.

Halibut in the commercial landings range from 7 years old and 4.5 kg to over 30 years old weighing over 100 kg. Commercial landings from the southeastern Bering Sea reached a peak of more than 7,000 mt in 1963. Current landings are about 250 mt. Historically management of the halibut resource has been the responsibility of the International Pacific Halibut Commission. Between 1963 and 1976 management was assumed by the International North Pacific Fisheries Commission, with recommendation from the International Pacific Halibut Commission to the Canadian and United States governments. With the advent of the Fisheries Conservation and Management Act of 1976, management responsibility was returned to the International Pacific Halibut Commission subject to approval by Canada and the United States.

INTRODUCTION

The Pacific halibut, *Hippoglossus stenolepis* Schmidt, the largest member of the family of flounders called Pleuronectidae, occurs in waters of the North Pacific Ocean and Bering Sea. This large, commercially valuable flounder has been the object of an intense commercial fishery in the northeastern Pacific Ocean since 1888 and was used for subsistence before that. A journal of Captain Cook's voyage of exploration reports the use of "holybret" by the crews of his vessels, natives, and Russian fur traders at Unalaska in 1777 (Munford 1963).

An intensive commercial fishery in the early 1900's off the continental United States and British Columbia and in the Gulf of Alaska depleted the resource and created the political climate which brought into existence the convention between the United States

and Canada "For the Preservation of the Halibut Fishery of the Northern Pacific Ocean Including Bering Sea," which was ratified 21 October 1924. This original convention provided for the establishment of the International Fisheries Commission (IFC), renamed International Pacific Halibut Commission (IPHC) in 1953, to provide for the management of the halibut resource. This convention, which was amended in 1930, 1937, 1953, and 1979,¹ still provides the means of managing the halibut fishery on behalf of Canadian and United States citizens. During this 55-year period there have been some trying times, particularly for the management regime established for the Bering Sea.

The International Convention for the High Seas Fisheries of the North Pacific Ocean between Canada, Japan, and the United States came into force in 1953. This convention created the International North Pacific Fisheries Commission (INPFC) to manage the stocks of fish within the convention area, including the Bering Sea. A notable feature of this treaty was the principle of abstaining from fishing certain fully exploited stocks of fish. Under this provision Japan was required to abstain from fishing halibut in waters of the North Pacific Ocean and Bering Sea east of 175°W provided that stocks of that species continued to meet the qualifications requiring abstention. No determination regarding abstention could be made until the convention had been in effect for five years (Bell 1969).

From 1958 through 1961, INPFC annually reviewed the requirements of the halibut stocks for continued abstention. In 1962, it was agreed by the three contracting parties that halibut in the Bering Sea east of 175°W no longer qualified and halibut in that area was removed from the abstention list. Japanese nationals were permitted to fish halibut beginning in 1963. Other countries not signatory to

¹The 1979 amendment had not been ratified by either Canada or the United States at the time of writing.

these conventions were free to fish halibut in the Bering Sea without limitation outside the fishing zone established by the United States.

With the enactment of the Fisheries and Conservation Management Act of 1976 (FCMA), the control of fishing within 200 miles of the coast of the United States became the exclusive concern of the United States Government. But recommendations to the government for the management of the halibut fishery in the Bering Sea are still the responsibility of IPHC.

The development of a halibut fishery in the Bering Sea was inhibited by the distance from established processing plants and transportation facilities and the presence of fishing grounds closer to the home ports of the fishing fleet. The relatively large landings from the Gulf of Alaska have dictated that much of IPHC's research activity be directed to that area to fulfill management directives.

LIFE HISTORY

Samples of all stages in the life history of halibut from larvae to mature adults have been taken from the waters of eastern Bering Sea. The complex process of evaluating the contribution of each stage to the ecosystem has not yet been accomplished.

Description

Early ichthyologists describing fish of the eastern Pacific Ocean believed the halibut there to be identical to the Atlantic halibut, *Hippoglossus hippoglossus*. Schmidt (1904, 1930) noted some differences in the shape of the scales, length of the pectoral fin, and shape of the body and described it as a separate species, *H. stenolepis*. Vernidub (1936) in further investigations did not consider these differences to warrant specific status and considered the Pacific form to be a sub-species, *H. h. stenolepis*; Schmidt (1950) later agreed with Vernidub. Since that time North American ichthyologists have detected other differences great enough to separate species and at the present time the American Fisheries Society (1970) accepts *H. stenolepis*.

Halibut are compressed laterally and, except in the larval stages, have both eyes on one side of the head. Adults can be distinguished from other North Pacific flounders by their large size; large, almost symmetrical mouth with conical teeth; a pronounced arch of the lateral line above the pectoral fin; and the crescent-shaped or lunate tail. Halibut are usually dextral, that is, both eyes are on the right or colored side, which is directed toward the surface. Rare specimens have the eyes and color on the left side of the fish.

Color varies from olive to dark brown or black, with lighter, irregular blotches. The color pattern of the ocean floor often influences the intensity of the color. The left or blind side is white, and faces the ocean bottom. Halibut are thinner than most other flatfishes, averaging about one-third as wide as they are long. The small scales are well buried in the skin, giving the fish a smooth look.

Distribution

In the eastern Pacific Ocean halibut are found from northern California into the Bering Sea. The distribution continues across the Bering Sea and off the Asiatic Coast between Hokkaido Island and the Gulf of Anadyr, as well as in the Sea of Okhotsk. Presently the largest concentrations are in the Gulf of Alaska, with a smaller population in the Bering Sea.

Halibut are demersal, living on or near the bottom, usually in water of 2-8 C, although they will tolerate colder temperatures. Younger ages are found in relatively shallow inshore areas while adults have been reported as deep as 1,000 m and adult halibut are regularly taken at depths of 300-500 m in the Bering Sea during the spring fishing season.

In trawl hauls made in the eastern Bering Sea, the youngest halibut have been found in the southern part but larger, older juveniles and adults are known to range into the more northern part. The younger fish may not have the strength and endurance to make the longer migration and compete with the larger individuals even though the environment is suitable during the summer.

The farthest north that halibut have been taken during IPHC surveys was 62°30'N. Two juveniles were caught in three hauls at this latitude during July. September surveys made by National Marine Fisheries Service (NMFS), when the waters are near their warmest temperatures, have reported a "small" halibut at 65°15'N, and a single adult halibut at 66°02'N, 168°02'W (Best 1977).

Seasonal movements

During the winter months, ice covers much of the area and water temperatures near bottom drop to 0 C or less; this condition forces the halibut to concentrate in the deeper, warmer water along the continental edge. With the advent of spring the ice retreats and the water gradually warms to temperatures that permit the halibut to disperse over the expanse of shallow flats, which provide a suitable environment for a nursery for young halibut as well as feeding grounds for the larger juveniles and adults.

Juvenile halibut were abundant at depths of 330-370 m between Unimak Pass and the Pribilof Islands during an IPHC survey in March; few were taken at shallower depths. Novikov (1964) reported concentrations of halibut in the same area in March. In April a concentration of halibut was reported near the northern entrance of Unimak Pass (ca. $54^{\circ}40'N$, $165^{\circ}09'W$) at depths of 80 and 104 m where bottom temperatures were 3.1 and 3.8 C. On the same cruise east of $164^{\circ}W$, when bottom temperatures were -1.0 to 1.4 C, no halibut were taken (Best 1977).

As the warming progresses, young halibut move eastward along the north side of the Alaska Peninsula and usually are found throughout Bristol Bay in June. By late June, they spread northward toward Nunivak Island. An IPHC survey in the vicinity of St. Matthew Island ($60^{\circ}30'N$) during the latter part of June found bottom temperatures to be 0 C or less and only 10 halibut were caught in 14 hauls. Temperatures of 4 C or greater were encountered south of $59^{\circ}30'N$ and catches increased to 86 fish in five hauls. An IPHC charter found a concentration of adults in the vicinity of St. Matthew Island in August and September, but they left the area by mid-October. The occurrence of small halibut near shore in Norton Sound during July, August, and September was reported by Turner (1886).

Many of the tagged halibut released in the Bering Sea have been recovered within the area, confirming the annual migration from deep areas in winter to shallow areas in summer and return to deep water in the winter. Novikov (1970) reported that a halibut tagged in deep water north of Unalaska Island in April was recovered 32 days later on the flats. The fish had travelled 184 km, an average of 5.7 km per day.

IPHC tagged nearly 3,600 halibut near St. Matthew Island during August and September and 221 were subsequently returned. Although most (114) were recovered near the release site, 56 were recaptured from deep water along the continental slope between Unimak Pass and the Pribilof Islands, and west to 180° . Most of the recoveries from the deeper areas were made by the North American setline fleet during the April fishing season, although some were from the Japanese trawl fleet during the winter. An additional 51 recoveries were made in the Gulf of Alaska. Several dense concentrations of adult halibut, such as the above, have been found. These concentrations have provided some excellent short-term fishing, but after the initial exploitation the concentrations no longer exist. A schematic diagram of the winter and summer distribution of halibut in the Bering Sea is given in Fig. 31-1.

Figure 31-1a, b, and c. Generalized pattern of halibut distribution in the eastern Bering Sea.

Environmental Factors

Water temperature recorded near bottom during the IPHC summer surveys in the Bering Sea ranged from a low of -1.7°C at a station near mid-shelf in 1975 to a high of 10.5°C at a shallow station in Bristol Bay in 1965. During the 1960's, IPHC surveys caught few halibut at stations with temperatures of less than 2°C and catches were largest when the temperature was $4-5^{\circ}\text{C}$. Soviet research trawlers in the southeastern part reported the highest catch per hour at water temperatures of $3.5-5.5^{\circ}\text{C}$ (Novikov 1964). The Fisheries Agency of Japan (1972) reported increased catches of juvenile halibut when bottom temperatures were about 5°C . Commercial halibut fishing in the Gulf of Alaska is generally conducted in water of $3-8^{\circ}\text{C}$ (Thompson and Van Cleve 1936).

A period of cooler water temperatures occurred between 1970 and 1976. Some of the IPHC stations at which halibut had been taken in previous years could not be sampled during June of 1972 and 1975 because of drifting ice. No halibut were taken at several ice-free stations in the survey area where water temperatures were 0°C or lower. A comparison of bottom water temperatures during June of 1967 and 1972 (Figs. 31-2A, B) indicates that under the cooler conditions halibut were found to be concentrated in a smaller portion of the southeastern Bering Sea than in warmer years. The limiting effect of the 2°C isotherm was less noticeable in the colder years. Warm water conditions returned to the Bering Sea again in 1977 and 1978 and juvenile halibut were distributed throughout the survey area much as they were in 1967.

The timing and extent of the annual movement from deep to shallow areas seem to depend upon the severity of the preceding winter. If the winter is mild and warming occurs early, the suitable temperature range will extend farther north. These conditions permit the halibut to migrate from the edge grounds onto the flats earlier in the year and open a large portion of the Bering Sea for feeding during the summer. If conditions are favorable, halibut may disperse as far north as Bering Strait. After a cold winter, warming is delayed and the area suitable for halibut is restricted to the more southern portion of the Bering Sea. When conditions inhibit northward dispersion, halibut tend to form more concentrated schools in the southern portion of the Bering Sea.

Subpopulation structure

The tagging studies conducted by IPHC, Japan, and the U.S.S.R. have shown some indication of the

movement of stocks between the eastern and western Bering Sea as well as into the Gulf of Alaska. The most notable movement was made by a fish released 5 July 1975 at $61^{\circ}18'\text{N}$, $175^{\circ}12'\text{E}$ during a joint IPHC-U.S.S.R. research cruise off Kamchatka, and recovered 19 May 1977 near the Shumagin Islands in the Gulf of Alaska. The rate of the interchange has not been calculated due to the small number of tags recovered and the lack of effective fishing effort on adults in the western areas.

Migration from the Bering Sea into the Gulf of Alaska, documented during the first tagging experiment by IPHC in 1930, has been confirmed by all tagging studies since. Many of the recoveries of tagged halibut released in the Bering Sea were made off southeastern Alaska and British Columbia, and as far south as northern California (Fig. 31-3). One of the longer movements was from St. Matthew Island to Hecate Strait, British Columbia, a distance of over 3,200 km; the fish was captured within one year after tagging. This is an average movement of about 8.8 km per day.

Dunlop et al. (1964), after making corrections for fishing effort, calculated that 24 percent of the halibut tagged in the Bering Sea in 1959 emigrated to the Gulf of Alaska. Best (1977) reported that three tagged juveniles released in the southeastern Bering Sea were recovered by the setline fishery in the Gulf of Alaska.

The foreign trawl fleet made substantial recoveries from a group of tagged juveniles (mean length about 30 cm) released by IPHC near Unimak Island. These fish were tagged and released in July at $54^{\circ}45'\text{N}$, $164^{\circ}45'\text{W}$ in depths of less than 50 m and were recovered from an area centered about $54^{\circ}50'\text{N}$, $165^{\circ}30'\text{W}$ at depths in excess of 200 m, a distance of about 50 km from the release site, the following winter. Two longer-term recoveries from this release have been recorded, one from west of the Pribilof Islands ($58^{\circ}35'\text{N}$, $177^{\circ}18'\text{W}$) and the other from the Aleutian Chain near Amukta Island ($52^{\circ}50'\text{N}$, $171^{\circ}25'\text{W}$).

If the theory of compensatory emigration according to which young halibut migrate to "counteract the drift of the natant eggs and larvae and to maintain the species in its habitat" (Dunlop et al. 1964) is valid, the two longer-term tag recoveries would suggest that some of the juveniles in the southeastern Bering Sea originate from spawning grounds to the west, while the three recoveries from the Gulf of Alaska suggest spawning to the east. Certainly these few tag recoveries are insufficient to reach any conclusion on migratory patterns, but they do make for some interesting speculation. A logical explanation is that the shallow areas of the southeastern

Figure 31-2a. Distribution of halibut in number per 60-minute haul and approximate location of bottom isotherms (C) in June 1967.

Figure 31-2b. Distribution of halibut in number per 60-minute haul and approximate location of bottom isotherms (C) in June 1972.

Figure 31-3. Recoveries of halibut tagged in the Bering Sea during 1959. The number of fish tagged is shown in the black box.

Bering Sea may be a nursery area for more than one stock of halibut.

Two tagged juveniles released on the Pacific Ocean side of Unimak Island at approximately the same time as the above Bering Sea releases were recovered in the Bering Sea the following winter by the same trawl fleet. This movement of two small fish was only about 125 km, but it is the only record of any movement of halibut from the Pacific Ocean to the Bering Sea. These scattered fragments of information only open the door to speculation and do not define the interchange of halibut throughout the Bering Sea and the adjacent areas.

Maturity

When all information on maturity collected from research cruises in the Bering Sea east of 175°W was taken into account, the age at 50 percent maturity for female halibut was calculated to be 13.8 years, at a length of 122 cm. Males averaged 7.5 years and 72 cm. This is older than the age (approximately 11 years) but about the same size reported for female halibut from the Gulf of Alaska (Schmitt and Skud 1978). In the Bering Sea fewer than 1 percent of the nine-year-olds and about 1 percent of the 90-cm females were observed to be mature. These data fit well with the information reported by Novikov (1964) for the eastern Bering Sea.

Novikov (1964) also reported an average fecundity of 1,164,000 eggs for 12 females between 100 and 120 cm in length and 2,338,000 for females between 120 and 140 cm from the Bering Sea during the period 1957-60. IPHC has no information on fecundity of halibut from this area. However, a 1973 study in the Gulf of Alaska (Schmitt and Skud 1978) reported only about half as many eggs for fish of the same size. The increased fecundity may be the result of an evolutionary process to ensure survival in the environment of the Bering Sea; or there may be a discrepancy in the methods of calculation used in the two studies.

Spawning

Studies conducted in the Gulf of Alaska have provided information on halibut spawning (Thompson and Van Cleve 1936). Information to date indicates that halibut spawning in the Bering Sea is similar in nature. An IPHC research cruise during the winter of 1963-64 found mature unspawned fish at depths of 250-550 m between Unimak Island and the Pribilof Islands in December. The same general area was fished again in January and most of the fish were spawned out. Spawning undoubtedly takes place at other locations along the continental edge west of the Pribilof Islands as well as along the Aleutian Islands.

The temperature of water at that depth remains a fairly constant 3-4 C the year round; consequently the developing halibut ova probably experience nearly the same conditions as in the Gulf of Alaska. Van Cleve and Seymour (1953) estimated that hatching would take about 23 days at 4.7 C. Under controlled laboratory conditions Forrester and Alderdice (1973) observed hatching after 20 days at 5 C but at 4 C ova did not survive to hatching. Under conditions found in the Bering Sea hatching would probably require at least three or four weeks.

The accepted pattern of circulation in the Bering Sea indicates that any spawning southeast of the Pribilof Islands should be carried in a northwesterly direction along the continental slope to the Asiatic Coast (Favorite 1974). Larvae have been captured west of the Pribilof Islands (see Waldron, this volume), providing a measure of circumstantial evidence to support this theory, and fish-of-the-year have been taken from shallow areas of the western Bering Sea. Halibut larvae also have been collected from locations near Unimak Pass (see Waldron, this volume). These collections suggest either that spawning takes place along the Aleutian Islands or that the larvae drift through the Aleutian Island passes from the Gulf of Alaska.

Thompson and Van Cleve (1936) reported newly hatched larvae at greater depths than the eggs in the Gulf of Alaska. However, by the age of three to five months all larvae were in the upper 100 m. Young halibut completed metamorphosis and were on the bottom by May or June at a length of 22-29 mm. Musienko (1957) reported taking young-of-the-year halibut 34-42 mm long in September and October off the Kamchatka Peninsula. This period of time for development through metamorphosis seems reasonable for the temperature regime found in the Bering Sea.

Although the early larval development takes place in rather deep water, the growing larvae gradually rise towards the surface waters. If conditions of currents and food are satisfactory the young halibut should find itself in relatively shallow water close to shore when metamorphosis occurs and the larvae change from a pelagic to a bottom habitat. Although metamorphosed young-of-the-year have not been reported from the eastern Bering Sea, they probably exist; IPHC regularly captures one-year-old halibut during late June from water less than 15 m deep along the northern shore of the Alaska Peninsula and from the bays on the Bering Sea side of the Aleutian Islands, using special small-mesh nets.

Size and age composition

The size and age of halibut captured depend to a large extent upon the type of fishing gear used. The juvenile halibut survey by IPHC and fishery resource surveys by the NMFS are conducted with a standardized trawl net which catches small halibut mostly in the range of 25-60 cm. The minimum size limit for commercially caught halibut is 81 cm, which is about equal to 4.5 kg. Setline gear will catch some small halibut but the undersized fish are discarded at sea. The commercial landings sampled by IPHC contain halibut from 80 to more than 200 cm. Some fish as young as seven years old will begin entering the setline fishery (a year-class is not fully recruited until age 10) and a few fish are caught each year that are over 30 years old and weigh more than 100 kg. The size distribution and age composition from the juvenile survey and commercial landings taken east of 175°W in 1977 are shown in Fig. 31-4.

Figure 31-4. Length and age distribution of halibut from survey trawl fishing and commercial setline landings.

A curious phenomenon of alternating strong and weak year-classes was reported from the juvenile surveys (Best 1977). The pattern of relative year-class strength generally has continued into the older ages in the 1977 commercial landings, where ages 10, 12, 14, and 16 have maintained greater abundance than the intervening ages (see Fig. 31-4). The 1961 year-class, which produced the largest number of 2-year-olds recorded in the juvenile surveys, was still above average as 16-year-olds.

Growth

Although female halibut are larger than males at any given age, the difference is very small until about five or six years of age. After that time the difference becomes significant and continues throughout the life of the fish. The rate of growth slows noticeably for males at about nine years of age and a length of 90 cm. The decrease in growth rate does not occur in females until age 13 and a size of about 125 cm. The size and age at which the decrease in growth rate occurs in the two sexes suggests that it may be tied to the onset of maturity. A comparison of the length and weight of male and female halibut is given in Table 31-1. These data are actual measurements of fish examined on IPHC research or observer cruises during the period 1961-78.

A comparison of growth rates for halibut from different areas shows that fish in the southeastern

Bering Sea grow at a faster rate than halibut from the western Bering Sea (IPHC 1976); but according to Southward (1967) they do not grow as fast as fish in the Gulf of Alaska. He also reported an increase in growth of Bering Sea halibut beginning during the mid-1950's that apparently has continued until the present. Best (1977) reported that a series of cold years (1972-75) in the eastern Bering Sea had reduced the growth of juveniles. The year-classes affected by the reduced growth are only now entering the commercial fishery and the short-term reduction in growth would likely be masked by the long-term average used in calculating the size at age.

Food habits

Since halibut are opportunistic feeders, utilizing whatever is readily available, a wide variety of food items has been found in halibut stomachs. Generally, the small halibut feed on small crustaceans; as they increase in size they begin to eat larger prey, including more fish. Little information is available on the food of adult halibut.

Examination of the stomachs of juvenile halibut from the Bering Sea indicated that they were feeding heavily on shrimp, hermit crabs, small shore crabs, and sand lance (Table 31-2). Novikov (1964) reported that halibut over 60 cm feed extensively on pollock and yellowfin sole.

TABLE 31-1

Average length in cm and round weight in kg at age for halibut caught east of 175°W

Age	Males		Females		Age	Males		Females	
	cm	kg	cm	kg		cm	kg	cm	kg
1	11	.01	12	.01	16	92	9.62	133	31.76
2	20	.07	20	.07	17	94	10.32	137	34.97
3	29	.23	30	.25	18	105	14.77	142	39.27
4	37	.50	39	.60	19	100	12.61	147	43.93
5	45	.95	49	1.25	20	98	11.81	147	43.93
6	56	1.93	63	2.82	21	111	17.68	154	51.05
7	67	3.44	78	5.64	22	102	13.44	161	58.99
8	77	5.41	87	8.03	23	99	12.20	158	55.50
9	86	7.73	97	11.42	24	107	15.70	164	62.63
10	89	8.64	103	13.88	25	106	15.23	167	66.41
11	98	11.81	111	17.68	26	111	17.68	165	63.87
12	98	11.81	118	21.56	27	—	—	178	81.66
13	98	11.81	128	28.06	28	—	—	182	87.76
14	95	10.68	131	30.24	29	—	—	164	62.63
15	89	8.64	131	30.24	>29	—	—	197	113.43

TABLE 31-2

Food items observed in halibut stomachs, southeastern Bering Sea, 1976-1977.
Numbers of stomachs in which each item was observed.

Food item	Fish length in cm								Total
	1-10	11-20	21-30	31-40	41-50	51-60	61-70	71-80	
Shrimp	9	47	36	7		2	1		102
Sand lance	1	11	12	7	4			1	36
Hermit crab		6	12	6					24
Shore crab			6	4	2	1			13
Isopods		7							7
Tanner crab		1			2			1	4
Pollock			2	1	1				4
Cottids		2					1		3
Smelt			2						2
Cod			1	1					2
Annelid worms		1							1
Euphausiids		1							1
Blenny		1							1
Sandfish							1		1
Sea poacher							1		1

In some areas of the Gulf of Alaska (although not significantly in the Bering Sea), halibut were found to feed heavily on Tanner crab. Halibut also have been found occasionally to feed on large king crab, particularly during the soft-shell stage (Gray 1964). In earlier years when the population of halibut was larger it is possible that halibut played some part in keeping the king and Tanner crab populations at lower than present levels of abundance. Reduced predation may have been partially responsible for the increase in the abundance of crabs.

Some predation upon halibut must take place, particularly at the smaller sizes, although there is little documented evidence. IPHC has, in the course of the food habit studies, found rare cases of cannibalism upon young halibut. By the age of five or six years halibut are one of the dominant predatory fishes and as such must be immune from predation except by the larger marine mammals. Halibut hooked on setline gear are easy prey for sea lions, which can render a portion of a catch unmarketable.

STOCK BIOMASS

The limited and sporadic nature of the commercial fishery has precluded any reasonable estimate of the biomass of adult halibut. However, IPHC was able to calculate a maximum sustained yield from the southeastern sector (between Unimak Island and

the Pribilofs) of 2,268 mt from data available between 1958-63 (Dunlop et al. 1964). This yield was exceeded in 1962 and 1963, and abundance, as indicated by catch-per-unit-effort, dropped sharply. With the added mortality of juveniles by the trawl fleets in southeastern Bering Sea, the fishery has never again reached that level.

Trawl surveys conducted by NMFS in 1975 and 1976 each produced estimates of nearly 31,000 mt of juvenile halibut (Bakkala and Smith 1978). The addition of adults, not available to the trawls used in this survey, would substantially increase this figure.

The removal of small halibut as incidental catch of the foreign trawl fleets during the 1960's and early 1970's has reduced the potential for recruitment and rebuilding of the adult halibut stocks. Time and area closures on the foreign trawl fleets have reduced the incidental catch of halibut in recent years. Although the FCMA will provide for even greater control of fishing in the southeastern Bering Sea, the halibut stock probably will never return to a size that will permit catches as large as those of 1962 and 1963, which were made from a virtually unfished stock.

The strength of year-classes observed as juveniles has fluctuated in alternating years through the 1960's. Recent surveys indicate that the 1973 year-class is large, but this group of fish will not become available to the setline fishery until 1981. A single large year-class cannot significantly improve the size of a population which is ordinarily made up of

over 20 year-classes. Several above-average years, preferably successive, will be needed to improve noticeably the supply of halibut.

HISTORY OF COMMERCIAL UTILIZATION AND REGULATION

The development of a fishery for halibut in the eastern Bering Sea was slow. The lack of processing facilities, the great distance from home ports, often adverse weather conditions, small stock size, and profitable fishing in the Gulf of Alaska all combined to keep fishing effort in the Bering Sea at low levels.

Fishery

A few U.S. vessels conducted a small fishery in the southeastern Bering Sea between 1930 and 1934. In those years the Bering Sea was open to fishing at the same time as the Gulf of Alaska. No further fishing occurred in the area until 1952, when U.S. vessels began taking about 100 mt annually from fishing grounds east of 175°W (Fig. 31-5). To encourage fishing in the Bering Sea, the fishing season

was opened one month earlier than in the Gulf of Alaska beginning in 1958. The catch began to increase, reaching nearly 4,400 mt in 1962 (Myhre et al. 1977), about equally divided between U.S. and Canadian vessels. INPFC determined that halibut in the Bering Sea no longer qualified for abstention and Japan was allowed to enter the fishery in 1963 (Forrester et al. 1978). Before this time, removals were limited only by the length of the fishing season, but in 1963 a three-nation catch limit of 5,000 mt was established by INPFC for that portion of the Bering Sea along the edge of the shelf between Unimak Island and the Pribilof Islands (roughly between 165 and 170°W), despite the fact that IPHC had calculated the maximum sustained yield from this 5° of longitude to be 2,268 mt for the period of 1958-63 (Dunlop et al. 1964). The 1963 catch from the quota area reached 4,974 mt and the total catch for the area east of 175°W was 7,254 mt. In 1964, a catch limit of 2,900 mt was set for the area between Unimak Island and the Pribilof Islands, but only 972 mt could be taken from the depleted stocks. Japan discontinued longlining for halibut after the 1964 season. Catches declined after that time, and since 1970 have averaged about 250 mt, largely caught by U.S. vessels.

Regulations

After the poor catches of the 1964 fishing season the open period was limited to only seven days in 1965 on the advice of IPHC. Since then the open period for the area east of 175°W has been gradually increased to its present 20 days in April and September, respectively. West of this line continuous fishing is permitted from April 10 to November 15. Poor catches in the eastern sector stimulated interest in fishing farther west, particularly along the Aleutian Islands.

Canadian interest in the Bering Sea fishery has declined with the growth of their domestic herring fishery, which occurs at approximately the same time as the spring season. The new "Protocol for Regulation of the North Pacific Halibut Fishery" eliminated Canadian participation in the Bering Sea, beginning in 1979.

IPHC adopted a minimum size for commercially caught halibut of 2.2 kg (5 pounds) dressed weight with head off in 1940. This minimum size was further specified in 1944 as 66 cm (26 in.) in length. The minimum size was increased to 81 cm (32 in.) in 1974, about equal to a 4.5-kg (10-pound) fish, dressed with the head off.

In addition to the directed setline fishery for adult halibut, very large numbers of young halibut have

Figure 31-5. Landings of halibut reported by Canadian, Japanese, and U.S. fishermen from the Bering Sea east of 175°W.

been taken incidentally in the fisheries for other species. Hoag and French (1976) estimated the incidental catch of halibut by foreign fleets from the entire Bering Sea at 11,519 mt in 1971. No estimate of the additional catch by the traps of the domestic crab fishery has been made. Incidental catches of mostly young fish have reduced the recruitment into the adult stocks. International negotiations initiated time-area closures beginning in 1974 for locations and times of greatest incidental catches of halibut. These restrictions have significantly reduced the incidental catches of small fish, which should improve recruitment to the adult stock in the future. The first indication of improvement has been noted in the increased numbers of juveniles in the annual surveys conducted by IPHC (Fig. 31-6).

HISTORY OF RESEARCH

The U.S. Bureau of Fisheries Steamer *Albatross* reported a few small halibut taken in the course of a survey of the codfish resource of the Bering Sea in 1890 (Rathbun 1894). The *Albatross* returned to the Bering Sea in 1911 in the course of a survey of the halibut grounds of the Pacific Coast (Alexander 1912). Only a few small halibut were caught off Akun and Akutan Islands, however; two sets with longline gear made near Unimak Pass caught no halibut and the vessel proceeded to the Pacific Ocean. The next research in this area was the tagging of setline-caught halibut in Makushin Bay off Unalaska Island by IFC in 1930. A small U.S. fishery began about this time and lasted for three or four years. With the rebuilding of the stocks in the Gulf of Alaska the fleet was able to secure profitable catches from grounds closer to port than the Bering Sea. The increased fishing in the Gulf of Alaska required that IPHC direct its research and management efforts to that area.

Research in the Bering Sea had a very low priority until funds were appropriated that permitted IPHC to initiate a large tagging program in the 1950's. In 1956 an IPHC-chartered setline vessel found substantial numbers of commercial-sized fish along the shelf-edge between Unimak and the Pribilof islands. The purpose of chartered trips was primarily to tag halibut, but fish unsuitable for tagging were a source of much life history information. The accumulated information was summarized by Dunlop et al. (1964).

In addition to the investigations of the halibut stocks available to setline gear, IPHC began a trawl survey of the flats (continental shelf) in 1963 which has continued to date. This study, primarily of the juvenile halibut in the area, has been summarized

Figure 31-6. Abundance of juvenile halibut (< 65 cm) in the southeastern Bering Sea.

through 1977 by Best (1977). The NMFS has also collected information on halibut numbers and distribution in conjunction with king crab and ground-fish surveys (Pereyra et al. 1976, Bakkala and Smith 1978). Japan also contributed to halibut research as part of its commitment to INPFC. Most of the data collected by Japan and IPHC contributed to the management decisions of the INPFC and are included in Pereyra et al. (1976). The U.S.S.R. made a comprehensive trawl survey between 1957 and 1961 and the information on halibut collected during this period was summarized by Novikov (1964).

A joint IPHC-U.S.S.R. research program to tag halibut began in 1975 with a cooperative cruise in the western Bering Sea by a U.S.S.R. vessel with IPHC and U.S.S.R. scientists on board (IPHC 1976). IPHC has continued releasing tags with legends in English, Japanese, and Russian since 1975. The agreement called for the U.S.S.R. to conduct a similar program in the western regions. Information gained from this cooperative research will provide a better assessment of the halibut resource.

The role that the Bering Sea plays in the life history and distribution of Pacific halibut, although of great importance, is not well understood. Looking at a map of Alaska, one first sees the Bering Sea as a large water mass isolated from the Pacific Ocean by the chain of the Aleutian Islands. In reality, the impression of isolation is false, for the many passes between the islands provide passage for Pacific Ocean

water, immigration routes for the pelagic eggs and larvae of halibut, and pathways for later emigration of adults.

RESEARCH RECOMMENDATIONS

The research carried out in the Bering Sea to date has been concentrated in the eastern part. Even the research efforts of Japan and the U.S.S.R. have been conducted in a disproportionate amount in the eastern part with little if any research reported for the western part.

There is a basic need to develop standardized catch and effort statistics on all fishing by all nations for the entire Bering Sea. Forrester et al. (1978) did an excellent job of compiling the available information through 1970.

A coordinated tagging program should be developed, stressing areas of the western Bering Sea and Aleutian Islands. Emphasis should be placed on tagging the juvenile halibut, which apparently make extensive movements from the nursery areas to the grounds where they become available to the commercial fisheries. Adults also need to be tagged to provide information on the seasonal exchanges of the commercial-sized halibut between winter spawning grounds and summer feeding grounds. The return of recaptured tags must be encouraged, in order to provide the greatest information from these efforts.

In conjunction with the tagging program much basic biological information on halibut can be collected from fish unfit for tagging. A simple but neglected project is the comparison of specimens from the eastern and western Bering Sea and the eastern and western Pacific Ocean areas by serological or electrophoretic techniques to determine population similarities or differences.

Any future research should be coordinated among scientists of all nations fishing in the Bering Sea. Communication among researchers is necessary for the best utilization of collected information. In a tag recovery program, communication with the actual fishing fleet is essential if much of the information is not to be lost or distorted before it reaches the issuing agency.

SUMMARY

Pacific halibut are found from northern California into the Bering Sea and across to the Asian coast. Their center of abundance is in the Gulf of Alaska, with a smaller but important population in the Bering Sea.

An inshore feeding migration in summer and an

offshore spawning migration in winter have been observed. The timing and extent of the inshore migration are influenced by environmental conditions, with cooler water tending to restrict the movement. A period of relatively cool water temperatures altered the summer feeding migration between 1970 and 1976. A directed movement of young fish, probably to counteract the drift during the prolonged egg and larval stages, has also been noted. Movement by tagged young halibut in both an easterly and westerly direction away from the tagging location suggests that there is a mixing of stocks in the Bering Sea. Based on returns of tagged fish, a 24-percent immigration into the Gulf of Alaska was calculated.

Female halibut in the Bering Sea mature at about 13 years of age and a size of 122 cm. A female will produce from one to two million eggs or more depending upon its size. Males mature at a smaller size of about 72 cm and at an age of about seven years.

Scientific observation indicates that spawning occurs at depths of 250-550 m in the area between Unimak Island and the Pribilof Islands between December and February. Capture of a few larval halibut suggest that spawning also occurs at locations west of the Pribilof Islands and along the Aleutian Islands. Halibut larvae have a pelagic life for at least five to six months and then settle in shallow water near shore.

Halibut in the commercial landings are from 7 to (occasionally) over 30 years old and range in size from 80 to over 200 cm. Female halibut are larger than males at each age with the difference becoming quite large at the older ages.

Examination of the stomachs of young halibut indicates a preference for shrimp, small crabs, and small fish, with larger fish utilizing larger prey items. The diet is varied; locally abundant prey of the proper size is heavily utilized.

Development of a commercial fishery for halibut in the Bering Sea was hampered by lack of facilities, distance from the home port, and adequate fishing in the Gulf of Alaska. Special seasons stimulated interest in the halibut of the area and landings slowly increased during the early 1950's, accelerated as the fleet developed a knowledge of the area, and reached a peak of over 7,000 mt in 1963. Catches have averaged about 250 mt during the 1970's. Rebuilding of the adult stocks has been hampered by the massive incidental catch of small halibut by the trawl fisheries for other species. Restrictions on trawling activities give promise of improved stock condition in the future.

Regulation of halibut fishing by U.S. and Canadian fishermen has been the responsibility of the IPHC since 1932. The INPFC managed the halibut resource in the Bering Sea from 1963 to 1976, while IPHC made recommendations to the Canadian and United States Governments for presentation to INPFC. After 1976, management of the fish resources of the area became the responsibility of the United States under the FCMA. However, the responsibility for halibut remains with IPHC, a joint Canadian-U.S. agency.

Research on halibut has been strongly influenced by the need for stock assessment and management of the resource. Between 1950 and the present a considerable amount of research has been conducted in the southeastern Bering Sea by IPHC and fishery agencies of the United States, Japan, and the U.S.S.R. This research has been largely uncoordinated and of an exploratory nature to develop and monitor the commercial fisheries in the area. Future research should be coordinated to provide a broad spectrum of information and to reduce costs of research by preventing duplication of effort.

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Distribution, Migration, and Status of Pacific Herring

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ABSTRACT

Pacific herring are an important part of the Bering Sea food web and form the basis of a major commercial fishery. Until recently Japan and the U.S.S.R. have been major exploiters of herring. Catch peaked in the early 1970's at 145,579 mt, and then declined in response to overfishing and poor recruitment. But recently herring abundance has increased, and the United States has become the dominant exploiter.

Most herring are harvested in coastal waters during the spawning period, which begins in late April/mid-May along the Alaska Peninsula and Bristol Bay and progressively later to the north. Spawning occurs at temperatures of 5-12 C and the time of spawning is related to winter water temperatures, i.e., early in warm years and late in cold years. During spawning, eggs are deposited on vegetation in the intertidal zone of shallow bays and rocky headlands. Eggs hatch in two to three weeks as planktonic larvae and metamorphose to juveniles after six to ten weeks. Little is known of larval and juvenile stages in the eastern Bering Sea.

Sexual maturity begins at age two, but most herring mature at ages three and four, the ages of recruitment to the fishery. Herring as old as 15 occur, but very few beyond age 10 are present in commercial catches. Age-specific mortalities are unknown but the average instantaneous natural mortality rate is estimated to be 0.46-0.47. The rate of growth is generally greater in the Bering Sea than in the Gulf of Alaska, but within the Bering Sea growth decreases to the north. Herring feed on the predominant larger zooplankton—euphausiids and copepods.

Three major stocks occur in the Bering Sea: northwest of the Pribilof Islands, the Gulf of Olyutorski, and Cape Navarin. These have been identified as individual stocks based on differences in growth and maturation rates, and dissimilar age structures. Herring wintering northwest of the Pribilof Islands migrate to the Alaska coast in spring and spawn in Bristol Bay and between the Yukon and Kuskokwim Rivers. Although some may also spawn in the eastern Aleutian Islands, Alaska Peninsula, and Norton Sound, herring in these areas may also winter inshore near spawning grounds. North of Norton Sound some herring winter in brackish lagoons and estuaries.

Herring wintering northwest of the Pribilof Islands arrive on the winter grounds in October and concentrate in waters of 2-4 C at depths of 105-137 m through the winter. In March schools migrate to the coast for spawning. After spawning they must remain in coastal waters, for few are found on the shelf or slope. In late August, they reappear in offshore waters in the areas of Unimak and Nunivak islands, and the seaward migration to the winter grounds continues through September and October.

Although assessments of eastern Bering Sea herring have ranged from 374 thousand mt to 2.75 million mt, the current estimate of spawning biomass is 432-864 thousand mt. Fisheries data indicate that herring declined rapidly after peak harvests in the early 1970's and that peak catches were supported by a few strong year-classes. Weak year-classes occurred through the early 1970's, and recruitment appears to be normalized in recent years.

Research is needed to refine estimates of abundance and biological characteristics of stocks, to improve the ability to predict changes in resource abundance, composition, and availability, and to identify the origin and distribution of herring in offshore areas.

INTRODUCTION

The Pacific herring (*Clupea harengus pallasii*) is a member of the family Clupeidae, which has global distribution and includes about 50 genera and 190 species found mostly in tropical and temperate waters (Svetovidov 1952). Herring occurring in the North Atlantic and North Pacific areas are similar in appearance, but differ mainly in number of vertebrae (55-57 vs. 52-55). Pacific and Atlantic species also differ biochemically, with significant differences

observed in the frequencies of eight genes (S. Grant, Northwest and Alaska Fisheries Center, Seattle, Washington, personal communication). Svetovidov (1952) believes that Pacific herring arrived from the Atlantic some time between the Pliocene and the "post-glacial 'transgression'" via the Asian Arctic. Pacific herring also differ from Atlantic herring in spawning and migrational behavior: the former are spring spawners, whereas the latter may be spring, winter, summer, or autumn spawners. Pacific herring spawn between the intertidal zone and about 20 m and deposit eggs on vegetation, whereas Atlantic herring spawn in deep water on a gravel bottom. Pacific herring generally remain near the spawning ground the year round and do not make extensive seasonal migrations as many Atlantic stocks do.

In the North Pacific Ocean, herring are distributed along the Asiatic and North American continental shelves (Fig. 32-1); in Asia they range from Taksi Bay, near the mouth of the Lena River, to the Yellow Sea (Andriyashev 1954), and in North America from Cape Bathurst in the Beaufort Sea to San Diego Bay, California (Hart 1973).

Herring are an important part of the eastern Bering

Sea food web. They are pelagic planktivores, highly adapted with large mouths and numerous fine gill rakes for efficient utilization of euphausiids, copepods, and other zooplankton. In turn, herring are important prey for marine mammals, birds, and roundfish. Mathematical simulations of the ecosystem in this area by Laevastu and Favorite (1978) indicate that annual total mortality amounts to one half of herring biomass production and that 95 percent of total mortality is by predation. Given this level of mortality, it is understandable that herring stocks exhibit strong fluctuations in abundance with apparently small changes in fishing or environmental factors.

The abundance of herring declined sharply in the early 1970's and only recently has an increase become apparent. Although several hypotheses could be advanced to explain the strong population fluctuations observed, data are insufficient to conclusively establish a cause. Since rapid, marked changes in abundance are expected to occur in the future, it will be necessary to be able to identify the causes and predict their occurrence and magnitude.

Present knowledge is rudimentary and inferences

Figure 32-1. Geographic range of Pacific herring (*Clupea harengus pallasii*) in the North Pacific Ocean.

about many phases of life history must be drawn from other more thoroughly studied populations. Research is needed on all aspects of herring biology, especially interspecies interactions and environmental effects on herring.

FISHERIES

Archeological excavations in the area indicate that net fisheries had been developed as early as 500 B.C. (Hemming et al. 1978), and subsistence fishing for herring continues today in many native villages, especially in villages between the Yukon and Kuskokwim rivers where alternative food resources (i.e., salmon, moose) are absent or in low abundance (Barton 1978). Commercial herring fisheries developed in the northern Bering Sea around the turn of the century. Marsh and Cobb (1910) reported that a small fishery in Grantley Harbor on the Seward Peninsula in about 1906 supplied salt herring to Nome. Before 1909, another small herring fishery developed in Golovnin Bay, Norton Sound (Rounsefell 1930). Although the Grantley Harbor fishery apparently was short-lived (the last reported catch was in 1917, when 300 barrels were packed), the Golovnin Bay fishery operated until 1941.

The first large-scale herring fishery began in 1928, when a purse-seine fleet fished at Unalaska. Salteries were established at Dutch Harbor in subsequent years, and the catch increased to a peak of 2,277 mt in 1932 (Barton 1978). Catches ranged between 1,000 and 2,000 mt until 1937, and thereafter declined until 1946, when the fishery ended. Lack of demand and accompanying low prices for cured herring were the principal reasons for the demise of this fishery (Wespestad 1978a). A herring fishery resumed in 1959, when Soviet exploratory trawlers found wintering concentrations along the continental slope northwest of the Pribilof Islands (Dudnik and Usoltsev 1964). During the first season, 10,000 mt were harvested (Fig. 32-2). Catches increased in later years as effort increased but declined sharply in 1965 and 1966 when herring could not be located and effort was greatly reduced.

Japanese vessels began fishing for herring in the late 1960's. A trawl fishery was established on the winter grounds from November to April and a gillnet fishery, which operated off the western Alaska spawning grounds, from April through June (Wespestad 1978b). In 1977, the area east of 168°W and north of 58°N was closed to foreign herring fishing to protect native subsistence fisheries, and in 1978 the 168°W closure line was extended to the Alaska Peninsula. This greatly limited the gillnet fishery; no

Figure 32-2. Eastern Bering Sea herring catch 1960-79.

Figure 32-3. Catch and catch per unit effort (CPUE) relationship for large Japanese stern trawlers in the eastern Bering Sea, 1967-74.

foreign gillnetting occurred in 1978 and only a very limited amount in 1979.

Catch and effort peaked in the late 1960's and early 1970's and then declined (Fig. 32-3). The peak catch occurred in 1970 when 145,579 mt were harvested (see Fig. 32-2). Catch dropped abruptly the next year, increased slightly in 1972, and then declined until 1976, when an increase occurred. In 1977, an allowable catch of about 21,000 mt was established by the U.S. when the 200-mile Fishery Conservation Zone was established. Herring harvests have been maintained at this level to the present.

U.S. herring fisheries resumed on a small scale in Norton Sound and northern Bristol Bay in the late

1960's for herring roe and herring roe-on-kelp for the Japanese market. Harvests were small, generally under 100 mt, until 1977, when the catch increased to 2,550 mt. The fishery expanded further in 1978, and the catch rose to 7,025 mt. In 1979 the catch increased again to approximately 12,000 mt. Most of the U.S. harvest is taken with purse seines and gillnets in northern Bristol Bay between Cape Constantine and Cape Newenham; smaller fisheries also occur in Goodnews Bay, Security Cove, Cape Romanzof, and Norton Sound (Fig. 32-4).

GENERAL BIOLOGY

Spawning

Herring spawn along the western Alaska coast in late spring to midsummer (Fig. 32-4). In most years, spawning occurs first along the Alaska Peninsula and in Bristol Bay in late April to mid-May and progressively later to the north. In Kotzebue Sound, it may extend from July through mid-August (Barton 1978). Spawning usually commences soon after the spawning grounds become ice-free; it has been noticed to begin

Figure 32-4. Distribution, time of spawning, and relative abundance of Pacific herring on coastal spawning grounds observed during aerial surveys, 1976-78.

when water temperatures are approximately 3-5.5 C (Scattergood et al. 1959), although it has been recorded over a range of 6-10 C in northern Bristol Bay (Warner and Shafford 1977) and in a range of 5.6-11.7 C on spawning grounds between Norton Sound and Bristol Bay (Barton 1979).

Prokhorov (1968) found that the approximate time of spawning in the western Bering Sea is related to winter and spring water temperatures with early maturation in warm years and delayed development in cold years. The past two years (1978 and 1979) have been mild winters, 1979 especially so, and herring have arrived at the spawning ground several days to two weeks earlier than average; in 1976, a cold year, spawning herring were not evident until mid-June. Svetovidov (1952) believes that the shore-spawning behavior of Pacific herring is caused by a lack of high water temperature in deeper water.

Spawning may last from a few days to several weeks. Generally, the older herring are the first to spawn, followed by successively younger fish. Eggs are deposited on vegetation in intertidal and shallow subtidal waters, predominantly on rockweed (*Fucus* spp.) and eelgrass (*Zostera* spp.) (Barton 1979). There are two types of spawning habitats: rocky headlands and shallow lagoons and bays (Barton 1978). South of Norton Sound most spawn is found in the intertidal zone on rockweed, while from Norton Sound northward most spawn is found in the shallow subtidal zone on eelgrass. Barton believes that these differences are partly due to smaller tide changes in the northern areas.

Eggs take 10-21 days to hatch, depending on water temperature. In northern Bristol Bay, hatching takes 13-14 days at 8-11 C (Barton 1979). However, Alderdice and Velsen (1971) have suggested that optimum temperatures for Pacific herring egg development are 5-9 C and that below 5 C eggs die.

Little is known of the magnitude or causes of egg mortality on eastern Bering Sea spawning grounds, but studies in British Columbia revealed wave action, exposure to air, and bird predation as major causes (Taylor 1964). Wave action may be a major cause in the Bering Sea; it has been observed that a severe storm after spawning activity destroyed both deposited eggs and rockweed in the upper intertidal zone along the south shore of Cape Romanzof (Gilmer 1978). Predation of fish may also be an important cause of egg mortality. Concentrations of flatfishes, particularly yellowfin sole, have been observed on the spawning grounds in northern Bristol Bay (John Clark, ADF&G, personal communication). Stomachs of flatfish examined in spawning areas by both authors have revealed a high rate of egg consumption.

Larval and juvenile development: General background information from Barkley Sound, B.C., studies

Herring hatch as larvae averaging 8 mm in size, and remain in this planktonic stage for approximately 6-10 weeks, at which time the larvae have grown to approximately 30 mm and begin to metamorphose into juveniles (Taylor 1964). During the larval stage, they are subject to high and variable mortality rates. An important reason for mortality may be failure to obtain proper food after the yolk sac is absorbed. Another may be passive transport away from the coast by prevailing currents (Outram and Humphreys 1974); Stevenson (1962) found that when larvae in Barkley Sound were transported to the open sea, few survived. He did not find temperatures and salinity to be important in inshore areas but believed that the high mortality of larvae offshore might be connected to the increased salinity of the open sea. There are indications that the direction and magnitude of surface stress may affect the survival of larvae; northward wind stress (along a N/S coast) will result in net onshore flow and water piling up against the coast, causing onshore retention of larvae and good year-classes. Offshore movement is associated with poor year-classes (Outram and Humphreys 1974).

When metamorphosis is complete, juveniles are free-swimming and begin to form schools that enlarge and move out of the bays as summer progresses (Taylor 1964). Hourston (1959) found that juveniles moved from the spawning grounds on the northwest side of Barkley Sound to rearing grounds on the southeast side. No specific reason could be found for migration to the southeast other than a preference for calmer, sheltered water found there. Juveniles in Barkley Sound actively feed at depths of 0.6-5 m at dawn and dusk. No sampling was done at night, but some inactive schools were observed near the surface. Juvenile schools were found in a range of salinities, but most were found at 25‰, which corresponds to Fujita and Kokudo's (1927) point of best fry survival.

Bering Sea

Little is known about the juvenile stage in the Bering-Chukchi Sea region from the time herring leave the coast in their first summer until they are recruited to the adult population. Rumyantsev and Darda (1970) indicate that juveniles feed in coastal waters in summer and move to deeper water in winter (juvenile herring in British Columbian and southern Alaska waters winter offshore and reappear in bays the following summer—Taylor 1964; Rounsefell 1930). In the western Bering Sea, herring aged 0

and 1 inhabit areas nearer shore and at lower temperatures than adults (Prokhorov 1968).

Of juveniles found in the Port Clarence area in 1977, more than 50 percent were captured in Imuruk Basin, the brackish forebay of the Port Clarence/Grantley Harbor complex. Apparently herring in the northern Bering Sea and Chukchi Sea may have a tolerance for much lower salinity, for eggs and fry were found in Imuruk Basin near Port Clarence in water of 40/00 (Barton 1978). Although juveniles were present in the spring spawning period (late June-early July), significant numbers were not captured until mid-August. Their presence in Hotham Inlet in November was indicated by stomach contents of sheefish (*Stenodus leucichthus*). Substantive numbers of age-one herring were captured in June 1978 in Hagemeister Strait of northern Bristol Bay (Barton 1979).

Wolotira et al. (1977) found both mature pre-spawning and immature herring in autumn (September-October) trawl catches made in the offshore waters of the northern Bering Sea and southern Chukchi Sea; however, age-0 herring were only found in the pelagic area of Norton Sound between Cape Douglas and Golovnin Bay.

Maturation and fecundity

Herring spawn for the first time at ages two to six but the majority do not spawn until ages three (50 percent mature) and four (78 percent mature). By age five, 95 percent of the population has matured (Rumyantsev and Darda 1970). Sexual maturity of eastern Bering Sea herring coincides with recruitment into the fishery, primarily at ages three and four. In the herring's southern range, the onset of sexual

maturity occurs earlier (stocks mature between ages three and four in British Columbia and ages two and three in California—Hart 1973, Rabin 1977).

In mature herring, fecundity increases with body length and latitude (Nagasaki 1958); fecundity appears also to be higher in the eastern Bering Sea area than in the Gulf of Alaska or western Bering Sea (Table 32-1).

Age and growth

Herring have been found to live as long as 15 years (Barton 1978), and generally occur in substantial numbers from ages three to six, but when strong year-classes occur, ages seven to ten may comprise a substantial portion of the catch.

Stocks grow at about the same rate as those in the Gulf of Alaska and British Columbia until ages three to four, but growth is greater in the Bering Sea for older fish, and they achieve a greater maximum length and weight than the more southern stocks (Fig. 32-5). Rounsefell (1930) reported many herring of 380 mm in the catch at Unalaska (compared to a maximum of 330 mm reported for British Columbia—Hart 1973). In more recent investigations, Rumyantsev and Darda (1970) and Warner (1976) have found Bering Sea herring of 340-345 mm. Barton (1978) found that size-at-age in spawning concentrations along the western Alaska coast from Norton Sound northward is significantly smaller than in stocks to the south (Fig. 32-5).

A general growth curve was derived for eastern Bering Sea herring by applying von Bertalanffy's equation to data reported by Shaboneev (1965) from the winter trawl fishery:

$$l_t = L_{\infty} (1 - e^{-K(t-t_0)})$$

TABLE 32-1

Fecundity (in thousands of eggs) of Pacific herring in different areas of the northern Pacific Ocean

Area	Age	4	5	6	7	8	Source
E. Bering Sea	1963	26.6	34.4	46.1	59.5	70.8	Shaboneev (1965)
	1964	26.6	32.1	52.4	53.5	77.8	Rumyantsev & Darda (1970)
Alaska Peninsula	1976	Mean: 26.4 Range: 12.6-84.8 Ages IV-VI					Warner (1976)
Karaginskii Bay	1963		26.4	30.1	37.4		Kachina ¹
W. Bering Sea	1964			39.2	43.3	50.6	Kachina
Vancouver	1955	19.9	23.8	29.6	38.2	30.4	Nagasaki (1958)

¹ Cited in Rumyantsev & Darda (1970)

LENGTH VS AGE OF PACIFIC HERRING

Figure 32-5. Size-at-age comparisons of Pacific herring from selected areas in the eastern Bering Sea and the northeastern Pacific Ocean.

The parameters of von Bertalanffy's equation are: L_{∞} (maximum length in mm) = 324.5, K (growth rate) = 0.35, and t_0 (age in years, fish was 0 length) = 0.0261.

Warner (1976) computed a von Bertalanffy curve for fish captured in trawl samples in Bristol Bay. His coefficients were $L_{\infty} = 299$, $K = 0.18$, and $t_0 = 2.10$. These estimates, although lower, do not differ significantly from Shaboneev's data, given the variances reported by Warner.

Mortality

Beyond the larval stage the rate of natural mortality decreases sharply and continues to decrease slightly until about age five, when it begins to increase from senility, disease, and spawning mortality (Fig. 32-6). The magnitude of mortality in the juvenile stage is actually unknown but inferences from other fish populations suggest that it is highest in years of high egg and larval survival because of intense intraspecific food competition (Ricker 1975).

A general estimate of the natural mortality rate for eastern Bering Sea herring was derived by Wespestad (1978a), using the procedure of Alverson and Carney (1975). This method estimates the natural mortality rate for eastern Bering Sea herring to be 0.47. Natural mortality can also be determined by an analysis of catch curves using regression techniques (Ricker 1975). Applying this method to data presented by Laevastu and Favorite (1977), we have determined the instantaneous mortality rate for fully recruited ages of eastern Bering Sea herring to be 0.46. Age-specific natural mortality estimates are unavailable for the eastern Bering Sea stocks; however, they are

TABLE 32-2

Instantaneous rate of natural mortality (M) for herring in the northeastern Pacific

Area/Age	3	4	5	6	7	8	Source
S. E. Alaska	.20	.30	.46	.59	.72	.85	Skud (1963)
E. Vancouver Is.			.40	.64	.77	.85	Tester (1955)
W. Vancouver Is.			.46	.61	.72	.79	Tester (1955)

probably similar to natural mortality rates estimated for herring stocks in southeastern Alaska and British Columbia (Table 32-2).

Food and feeding

The first food of herring larvae is usually limited to small and relatively immobile plankton organisms that the larvae must literally nearly run into to notice and capture. Microscopic eggs sometimes make up more than half of the earliest food; other items include diatoms and nauplii of small copepods.

Herring do not have a strong preference for certain food species but feed on the comparatively large organisms that predominate in the plankton of a given area (Kaganovskii 1955). Feeding generally occurs before spawning and intensifies afterward (Svetovidov 1952). During the winter, feeding declines; it ceases in late winter (Dudnik and Usoltsev 1964). During November and December, in Kamchatka waters of the western Bering Sea, Kachina and

Figure 32-6. Total mortality of Pacific herring expressed as percent annual cohort biomass loss.

Akimova (1972) found that juvenile herring consumed small and medium forms of zooplankton (chaetognaths, copepods, tunicates) and benthoplankton (mysids). Euphausiids, amphipods, mollusks, and other organisms were found rarely, and usually in small quantities. In the demersal zone, herring stomachs contained quantities of tubes of polychaete worms, bivalve mollusks, amphipods, copepods, juvenile fish, and detritus.

In the eastern Bering Sea, stomachs in August were 84 percent filled with euphausiids, 8 percent with fish fry, 6 percent with calanoids, and 2 percent with gammarids (Rumyantsev and Darda 1970). Fish fry, in order of importance, were walleye pollock, smelt, capelin, and sand lance. In spring, food was mainly *Themisto* (Amphipoda) and *Sagitta* (Chaetognatha). After spawning, the main diet was euphausiids, *Calanus* spp., and *Sagitta* spp. (Dudnik and Usoltsev 1964).

Nearly 75 percent of herring stomachs examined in the spring from Bristol Bay to Norton Sound either were empty or contained only traces of food (Barton 1978). Only 25 percent of the stomachs examined were at least 25 percent or more full, and only 3.4 percent were completely full. Major food items were cladocerans, flatworms (Platyhelminthes), copepods, and cirripeds.

DISTRIBUTION

Stock distribution

Three major herring wintering grounds have been identified within the Bering Sea: northwest of the Pribilof Islands, in the Gulf of Olyutorski (Prokhorov 1968), and near Cape Navarin (N. Fadeev, TINRO, Vladivostok, U.S.S.R., personal communication) (Fig. 32-7). Differences in the pattern of migration between the coast and the outer continental shelf have effectively isolated Asian and North American herring in the Bering Sea. The different growth and maturation rates and dissimilar age structures reported by Kachina (1978) of those wintering near Cape Navarin and those wintering northwest of the Pribilof Islands suggest that although these groups winter in close proximity there is little or no mixing between them. Most herring which winter near the Pribilof Islands are believed to spawn in Bristol Bay and in areas between the Yukon and Kuskokwim rivers. This conclusion is based on Soviet research, similarities in age composition, and the distribution of Japanese trawl catches during the spawning migration (Wespestad 1978b, Barton 1979). ADF&G aerial surveys indicate that the greatest abundance of spawning herring occurs in the Bristol Bay area

Figure 32-7. Range of Pacific herring in winter and spring.

and smaller spawning concentrations occur to the north and south (see Fig. 32-4).

The relation of herring spawning in Norton Sound to spawning stocks to the south is unclear. Those in Norton Sound are genetically similar to spawning stocks to the south (Grant 1979) and appear in inshore waters in late May to early June (Barton 1978), which suggests they may winter offshore. However, it is possible that some or all herring remain in Norton Sound the year round. Barton (1978) reported that an autumn, non-spawning run occurs in Golovnin Bay in northern Norton Sound, and herring have been caught through the ice by local residents jigging for cod in this area and near Nome. Moreover, herring have been found in ringed seal (*Phoca hispida*) stomachs collected near Nome in November. To the north of Norton Sound, herring occur in Port Clarence, and in inlets from the Bering strait to areas within Kotzebue Sound. Many, if not all, stocks found north of Nome remain in the immediate area the year round and winter in coastal lagoons and brackish bays, even though in several locations (e.g., Port Clarence, Shishmaref Inlet, and inner Kotzebue Sound) ice may cover the surface (Barton 1978).

Herring may also occur along the Alaska Peninsula and throughout the Aleutian Islands. Marsh and Cobb (1911) reported a large spawning at Atka Island in 1910, and spring and autumn runs (the latter presumably non-spawning) at Unalaska and Port Heiden. The fishery which operated at Unalaska in

Figure 32-8. Monthly distribution of Pacific herring by temperature and depth in the eastern Bering Sea. May-November data from Rummyantsev and Darda (1970) December data from Shaboneev (1965).

the 1930's and 1940's harvested herring in summer and early autumn, averaging 1,337 mt between 1929 and 1937. The current status of these stocks and their relationship to other eastern Bering Sea stocks are unknown. Recent aerial surveys by ADF&G have found small spawning concentrations on the north shore of Unimak Island, in Heredeen Bay, and in Port Heiden (Warner and Shafford 1977). Catches by Japanese trawlers just north of Unimak Pass in winter indicate that this may be the wintering area of herring spawning on the Alaska Peninsula (Wespestad 1978b).

Seasonal distribution—Pribilof stock

Temperature may influence seasonal distributions more than anything else. Soviet scientists found spring herring moving through sub-zero water temperatures on the way to spawning grounds, but during summer months they were found on the shelf in the warmer, upper layers of the water column (Fig. 32-8). As we said before, Svetovidov (1952) believes that migrations to the coast for spawning developed because of the lack of sufficiently warm water in spring and summer in the North Pacific Ocean. Furthermore, earlier warming of coastal waters provides earlier development of phytoplankton and zooplankton and hence better feeding conditions.

Winter

The major wintering ground of eastern Bering Sea herring is northwest of the Pribilof Islands, approximately between 57 and 59°N lat., representing an

area of 1,600-3,000 km² (Shaboneev 1965) that shifts in relation to the severity of the winter. In mild winters herring concentrate farther north and west, and in severe winters they move south and east (Fig. 32-7).

Dense schools are found during the day a few meters off the bottom at depths of 105-137 m and at water temperatures of 2-3.5 C (Dudnik and Usoltsev 1964). Very few were found shallower on the continental shelf where lower temperatures prevailed. Distinct diurnal vertical migrations occur in early winter; however, as the season progresses, diurnal movements diminish and herring remain on bottom during the day and slightly off bottom at night (Shaboneev 1965).

Spring

Soviet scientists investigating herring distribution in the mid-1960's found that herring left the wintering grounds in late March: they believed that herring followed two routes to the coast, and Japanese trawl catches in April and May indicate one major and one minor path (Fig. 32-7). The past two years (1978-79) have been mild winters, 1979 especially. In these years, herring arrived on spawning grounds along the coast several days to two weeks earlier than average; in 1976, a cold year, they were not found until mid-June.

Summer

The failure of Soviet surveys in the mid-1960's, using gillnets and trawls, to find herring concentrations on the Bering Sea slope or shelf in the summer suggests that most herring apparently remain temporarily in coastal waters after spawning. Annual NWAFC summer trawl surveys covering much of the continental shelf of the eastern Bering Sea support this conclusion, for very few herring have been taken in summer surveys (Pereyra et al. 1976, Bakala and Smith 1978). A hydroacoustic survey conducted along the outer shelf between Unimak Pass and the U.S.-U.S.S.R. convention line in June-July 1979 found only one herring in 2,558 nautical miles and 35 midwater trawl hauls. These results indicate that only a small number of herring may remain or return offshore in summer; most remain in coastal waters. Rummyantsev and Darda (1970) concluded that herring remained in coastal waters during the summer because heavy phytoplankton blooms (1-3 g/m³) and poor feeding conditions exist on the outer shelf. Herring captured on the outer shelf during the summer were in poor condition, perhaps because they had been feeding on items of low nutritional value—a diet other than their preferred

zooplankton. Other researchers have found herring to avoid areas of heavy bloom because of low nutritional value of phytoplankton and because the gill-clogging properties of certain phytoplankton species interfere with respiration (Henderson et al. 1936).

Concentrations began reappearing in offshore waters in the areas of Nunivak and Unimak islands in August (Rumyantsev and Darda 1970). The movement offshore in the area of Unimak Island appears to occur annually in August, for it is then that U.S. fishery observers on foreign vessels in that area first encounter herring in trawl catches in greater than trace amounts.

The distribution of herring between the time they leave the spawning grounds and the time they reappear in offshore waters is unknown, but salmon fishermen report catching large herring frequently in salmon gillnets in coastal areas of Bristol Bay in late June and July. Furthermore, Dudnik and Usoltsev (1964), using drift nets, found commercial quantities of herring only in littoral areas along the northern portion of the Alaska Peninsula. The reappearance of seaward migrants in late summer in two locations suggests a summer migration along the coast (Fig. 32-9). Migration to winter grounds continues through September with the herring progressively moving to deeper water and concentrating in the 2-4 C temperature stratum.

Fall

Concentration in the winter grounds begins in October and continues into winter. Mature fish were found to arrive at the wintering grounds before immature fish (Rumyantsev and Darda 1970). Immature fish had a tolerance or preference for colder, less saline waters of the shelf than adult fish (Fig. 32-10).

Seasonal distribution—north of Norton Sound

The annual cycle of herring to the north of Norton Sound appears to be markedly different from that in the central and southern Bering Sea. In these areas herring appear to move into brackish bays and estuaries for spawning and wintering, presumably finding temperatures rendered suitable by freshwater rivers and streams. Barton (1978) found herring spawning in Imuruk Basin, a brackish forebay of Port Clarence, in 4‰ salinity. The herring dispersed after spawning and reappeared in Imuruk Basin in mid-autumn.

Large numbers of herring were found concentrated just outside Kotzebue Sound in 6-8 C water in September and October (Wolotira et al. 1977). It is

Figure 32-9. Summer and autumn migration routes to winter grounds. Large solid arrow: area of reappearance in offshore waters as determined by Soviet research and Japanese catches. Large clear arrow: area of autumn reappearance in offshore waters reported from Soviet research. Small arrows: possible summer feeding routes and autumn migration routes.

possible that these herring move south with the advancing edge of the ice field, but some evidence suggests that they remain in the area through the winter. The phenomenon of herring moving into coastal waters in winter and offshore in summer has also been reported from Asia; Andriyashev (1954) reported that populations occur in Kamchatka, Sakhalin, and Honshu which winter and spawn in brackish lakes and lagoons. Barton (1978) cites ADF&G records of herring being found in sheefish (*Stenodus leucichthus*) stomachs which were collected in Hotham Inlet, Kotzebue Sound, in late November.

To the south of Kotzebue, on the northwest side of the Seward Peninsula, herring occurred in 11 of 14 stomachs of spotted seals (*Phoca largha*) collected at Shishmaref in October, and in 13 of 30 ringed seals collected in January and February. Marine mammal biologists who analyzed the stomachs indicated there was little doubt that herring were ingested in the area where seals were captured.

ABUNDANCE TRENDS AND STOCK STATUS

Estimates of absolute abundance are rare and even relative abundance data are rather limited. Attempts

Figure 32-10. Distribution of Pacific herring in October and the relationship of adults (mature) and juveniles (immature) to salinity gradients. (Modified from Rumyantsev and Darda 1970.)

have been made to estimate herring biomass by a Soviet hydroacoustic trawl survey, ecosystem modeling, and aerial surveys of spawning biomass.

In 1963, three years after the fishery began, the eastern Bering Sea herring biomass was estimated to be 2.16 million mt, based on a Soviet hydroacoustic survey of the wintering grounds (Shaboneev 1965). A recent paper by Kachina (1978), using the same data, reduced this earlier estimate to 0.374 million mt by using a lower mean school density—0.5 fish/m³—than the original estimate, which used 3.38 fish/m³. According to Shaboneev, schools were surveyed at night and the area and height of schools were mapped acoustically; school composition and age distribution were determined by trawling. The density used in the original estimate (3.38 fish/m³) was determined by comparing acoustic echograms from the eastern Bering Sea to echograms of schools sampled by purse seines in western Bering Sea coastal water. The revised estimate of 0.5 fish/m³ is based on densities observed in surveys of herring concentrations in the winter grounds northwest of the Pribilof Islands during 1969-71 (N. Fadeev, TINRO, Vladivostok, U.S.S.R., personal communication). The densities derived are questionable but cannot be fully evaluated because few specific details of Soviet survey methods and accuracy are available. However, data reported in the literature and from people involved with herring hydroacoustic surveys indicate that the

range of densities used by the Soviets may be extreme; an intermediate value may be more realistic.

Recently, a numerical ecosystem model was applied to estimate the biomass of eastern Bering Sea herring (Laevastu and Favorite 1978). This model based herring abundance on the amount needed to sustain the diet of herring predators at reported rates of consumption. The accuracy of input parameters, such as predator population size and consumption rates, has not been sufficiently evaluated, but model results show that a minimum stock size of 2.75 million mt of herring is required to maintain components of the ecosystem, including predators, at a level observed in the mid-1960's before the start of intensive fishing.

Aerial surveys have been flown in the past several years along the western Alaska coast during the spawning period to record the number of schools by surface area (Barton 1979). Estimates of biomass were obtained by converting estimated school surface area, using densities of 0.1 and 0.2 mt/m² of surface area.

The estimated spawning biomass along the coast from Bristol Bay to Norton Sound in 1978 was 432-864 thousand mt. These estimates include various errors—in determining surface areas and volumes of schools, in recording schools of other fish such as capelin, smelt, or sand lance as herring, and in recording the same school more than once during the season. They may therefore greatly overestimate actual spawning biomass (Barton 1979).

Biomass estimates are rather rudimentary, but CPUE data of the Japanese trawl fishery and ADF&G aerial surveys indicate that herring abundance declined sharply in the early 1970's and increased in the late 1970's. The CPUE (mt/hr) for Japanese large stern trawlers decreased from a high of 6.80 in 1969-70 to 0.77 in 1973-74 (Table 32-3). The CPUE of small stern trawlers also declined. The CPUE of the Japanese gillnet fishery exhibited no trend, presumably because vessels were targeting on spawning concentrations, which may not reflect population abundance (Wespestad 1978b).

ADF&G aerial surveys have indicated an increase in herring abundance in all major spawning areas during 1976-78 (Table 32-4). Preliminary assessment of observations in 1979 indicates an abundance similar to that of 1978, or slightly greater. The longest series of aerial counts, from southern Norton Sound, extends back to 1968; like the trawl CPUE data, it indicates a decrease in abundance during the early 1970's.

Because of changes in the fishery, the current level of herring abundance cannot be related to former

TABLE 32-3

Herring catch per unit effort data for the Japanese trawl and gillnet fisheries in the eastern Bering Sea.

Year	Stern trawl		Gillnet
	Small trawlers (mt/hour)	Large trawlers (mt/hour)	
1967	1.25	2.09	--
1968	1.75	4.63	.28
1969	0.81	6.80	.39
1970	1.06	6.74	.24
1971	0.56	1.52	.34
1972	--	1.84	.04
1973	--	0.77	.14
1974	0.29	0.17	.35
1975	--	--	.16
1976	--	--	--

levels; the catch and CPUE of foreign trawlers are no longer useful as indicators of abundance. Because of increased targeting on pollock, herring are not largely incidental catches to other fisheries, and the allowable catches have been low in recent years (Wespestad 1978b). Furthermore, the absence of aerial surveys in major spawning areas before 1976 precludes the use of this method for determining the relationship of past to present biomass levels.

Length and age frequency data indicate that catches in the late 1960's and early 1970's were composed of larger and older herring than in the past few years (Table 32-5). These data suggest that recruitment was poor until recently, a fact which may have contributed greatly to a lower herring abundance. Recruitment appears to have increased beginning with the 1972 year-class (Fig. 32-11).

Age-four herring comprised 54 percent of the catch in 1976, 50 percent in 1977, and in 1978, 65 percent of the purse seine catch. In 1979, preliminary results suggest that the recruitment of age-four herring has decreased from that observed in 1976-78; however, it appears that age-three fish are present in greater numbers than in the recent past.

Naumenko (1979) recently presented data showing several years of relatively weak year-class strength in the 1960's and early 1970's (Fig. 32-12). These data suggest that the peak catches of the fishery were sustained by a few strong year-classes and that future yields of this magnitude are likely only in the event of a series of much above average year-classes, or of a change in the ecosystem.

Figure 32-11. Age frequency of Pacific herring in the Bristol Bay herring roe fishery, 1976-78.

TABLE 32-4

Relative abundance indices of spawning herring standardized to 1976
in major spawning areas of the eastern Bering Sea.¹

	1968	1972	1974	1975	1976	1977	1978
Bristol Bay	--	--	--	--	1.0	2.1	20.0
Goodnews Bay/Security Cove	--	--	--	9.5	1.0	20.9	61.6
Nelson Island	--	--	--	0.5	1.0	1.0	3.2
Norton Sound:St. Michaels to Unalakleet	20.8	8.4	2.9	0.0 ²	1.0	--	4.2

¹ Relative abundance indices are corrected school counts weighted by surface area obtained from aerial surveys.

² Minimal survey effort.

RESEARCH REQUIREMENTS

Research is needed to refine estimates of abundance and define biological characteristics of stocks; to improve the ability to predict changes in resource abundance, composition, and availability; and to identify the origin and distribution of herring in offshore areas.

Estimates of biomass of specific groundfish resources have been obtained through resource surveys using bottom trawls; but since herring are not generally available to bottom trawls, other gear and methods must be used to assess biomass. Hydroacoustic surveys, spawn deposition surveys, and aerial surveys of schooled fish are under consideration for this purpose.

TABLE 32-5

Mean length of herring taken in the fisheries by all gear in all
months in the eastern Bering Sea and Alaska coastal water.¹

Year	Foreign trawl fishery			Coastal fishery		
	Mean Length (cm)	Sample size	Probable Average ages	Mean Length (cm)	Sample size	Location of sample
1964	26.60	3,101	7	23.30	339	Norton Sound
1965	29.83	155	8-9			
1966	27.16	48	6-7			
1967	26.20	99	5-6			
1968	29.04	4,771	8-9	28.60	350	Bristol Bay
1969	30.66	3,951	9-10			
1970	30.81	3,813	9-10			
1971	29.21	4,299	8-9			
1974						
1975						
1976			3-4	20.11	791	Bristol Bay
1977	23.40 ²	1,981	4-5	23.00	2,847	Bristol Bay
1978	23.28 ³	3,607	4-5-6	23.27	1,031	Bristol Bay

¹ Standard length for all coastal samples; fork length for foreign samples prior to 1978.

² Fork length (Nov. 1976-Feb. 1977) estimated standard length is 22.40 m.

³ Standard length (Dec. 1977-Jan. 1978).

Sources: Foreign fishery: Fisheries Agency of Japan, Rumyantsev and Darda (1970). U.S. observers on Japanese and Soviet vessels.
Coastal fishery: Alaska Dept. of Fish & Game, 1964 Annual Report; Bristol Bay Data Report No. 17; Barton et al. (1977). Warner & Shafford (1977).

Figure 32-12. Abundance of Pacific herring year-classes in the eastern Bering Sea relative to the 1975 year-class. (From Naumenko 1979).

Hydroacoustic surveys in the nearshore areas just before or during spawning are probably not practical because of the many widely scattered schools that are constantly moving through the shallow waters. Optimum results can be expected on the winter grounds, when herring are relatively stationary and concentrated. Results of surveys conducted during late winter to early spring could be applied in time for management of the roe fisheries.

Spawn surveys convert the amount of spawn deposited to size of adult population, using age-sex-size composition and fecundity data. Such surveys would have to be conducted immediately after spawning so as not to be affected by losses from predation and storms. The vast size of the area, including distances between spawning areas, lack of subtidal spawning information, and various logistical problems currently render this method impractical for the eastern Bering Sea.

In spite of limitations due to weather and narrow time-area coverage, aerial surveys may be one of the most cost-effective ways of measuring the abundance of spawning herring. School distribution within a limited area should be intensively studied to determine if surveys are more effective at particular times and to assess the variability of schools along sighting tracks. Aerial biomass estimation procedures and species identification procedures need to be developed. If a model of spawning school distribution could be developed, then statistical procedures could be used to overcome some of the weather and time limitations. Satellite technology may be used to augment aerial surveys—large schools may be observable at distances from the coast or spawn deposition (milt) may be observable. A combination of low-level aircraft and satellite observations may

provide solutions to the problems of effective coverage of tracklines and time-space distribution of schools.

Long-term fisheries management requires reliable forecasting of stock conditions. Until now, forecasts have been based mainly on past events, such as trends in abundance indices (CPUE) and size and age composition of specific resources, without any consideration of the interactions of these resources among themselves and with the environment. Studies need to be continued to determine, for predictive purposes, major influences on the abundance, composition, and distribution of resources. Monitoring certain oceanographic and climatological conditions (temperature, currents, etc.) in both the nearshore spawning and rearing grounds and the offshore wintering grounds may help us to understand fluctuations in herring abundance.

There is a critical need for annual measurements of the abundance of young fish before they enter the fisheries, in order to forecast later contribution to the exploitable stock. Assessment of pre-recruit abundance could be made of juveniles in nearshore nursery areas or at a later age in offshore waters. The major limitation of this method is the virtual absence of information about the distribution of eastern Bering Sea herring during the first two to three years of their life cycle.

Basic biological research is needed to systematically investigate population parameters, such as age-specific mortality rates, growth rates, and recruitment rates. Investigations are also needed to establish the degree of use of herring in the diet of marine mammals, salmon, and other predators so as to evaluate the ecological effects of harvesting.

Finally, stock distribution needs to be investigated so that individual stocks within the eastern Bering Sea can be monitored in relationship to other stocks and occurrence in fisheries.

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The Biology of Walleye Pollock

33

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ABSTRACT

Walleye pollock, *Theragra chalcogramma*, occur broadly distributed over the eastern Bering Sea outer continental shelf, and at their presently estimated population size of five million mt are a large source of organic production. Within the food web pollock are an important food resource for a wide variety of other fish species, marine mammals, and avifauna. Pollock also represent a major source of predation directed toward zooplankton and cannibalistic behavior. In addition, trawl fisheries harvest approximately 950,000 mt of pollock annually.

Although the genetic identity of eastern Bering Sea pollock is apparently distinct from populations in the western Pacific, genetic differentiation observed within the eastern Bering Sea is low, and the entire eastern Bering Sea population may be regarded as effectively a unit stock. The population is composed of approximately fifteen year-classes that show differences in geographical distribution and behavior based upon age. The overall rate of population mortality is high, approximately 51 percent dying annually, and age-class abundance is variable between years.

Spawning occurs predominantly along the southeastern Bering Sea outer shelf, just west and northwest of Unimak Island. Migratory movements that accompany ontogeny include northwest drift and broad dispersal by age one. In general, the adult fraction of the population undergoes seasonal movements to deep water in winter, to spawning sites in spring, then to the outer and central shelf in summer. Two apparent influences of ocean climate, warm or cold years, affect the timing and extent of these seasonal changes in geographical range.

There are fundamental needs for better understanding of important factors affecting the population dynamics of eastern Bering Sea pollock, to ensure collection of adequate time series data, and further studies of interspecific relationships and population exchange.

INTRODUCTION

The gadoid fish species walleye pollock, *Theragra chalcogramma* (Pallas 1811), is now recognized as perhaps one of the most important components of the Bering Sea biological system. Pollock have been found to represent a large fraction (ca. 20-50 percent: Pereyra et al. 1976) of the total standing stock of eastern Bering Sea demersal fishes, and the annual production of this organic pool is large. In 1978,

commercial trawl fisheries operating in the eastern Bering Sea harvested 977,700 mt of pollock (Pileggi and Thompson 1979). This amount represented 56 percent of all foreign fish catches in the fishery conservation zone (i.e., inside the 200-mile limit) of the United States, and a total dockside (ex-vessel) value of approximately \$92.4 million.

Because of their large standing stock, wide geographical distribution, and large rates of total food intake and production (growth), pollock represent an important part of the Bering Sea food web. During different life history stages pollock can be a major source of predation, feeding upon a broad spectrum of primarily zooplankton and fish prey. In turn, pollock themselves serve as an important food source for a variety of other demersal and pelagic fish species, pinnipeds, cetaceans, and avifauna.

POPULATION CHARACTERISTICS

Nomenclature

The walleye pollock, *T. chalcogramma*, is the only recognized member of the genus *Theragra* and is endemic to the North Pacific (Fig. 33-1). The overall species range extends continuously around the continental shelf areas of the northern Pacific Ocean and Bering Sea. In the northwest Pacific the species extends from the Sea of Japan, through the Okhotsk Sea and Kurile Islands, and along the Kamchatka Peninsula and Anadyr Gulf. In the northeast Pacific the range extends from St. Lawrence Island through the eastern Bering Sea, through the Gulf of Alaska, and along the northeastern Pacific coast to central California (Hart 1973). Pelagic populations have also been recently found over deep water in the central Bering Sea (Okada 1977, 1978).

Figure 33-1. The worldwide distribution of walleye pollock, *Theragra chalcogramma* (shaded area).

In the eastern Bering Sea, *T. chalcogramma* is by far the most abundant of the four common gadid species. These other species include *Gadus macrocephalus* (Pacific cod), *Boreogadus saida* (Arctic cod), and *Eleginus gracilis* (saffron cod).

Genetic structure

Although pollock are distributed (and perhaps spawn essentially continuously) along the northern rim of the Pacific Ocean and Bering Sea, genetic differentiation might reasonably be expected due to the isolating effects of both geographical distance and barriers (Kimura and Ohta 1971). This is because individual migrations will nearly always be considerably less than the total distributional range of the species, and instead, effectively form regional breeding populations. Similarly, diffusive exchange of eggs and larvae due to advective transport may be expected to be higher within regional current systems than between, although progressive transport of successive generations may be significant.

Consistent with these hypotheses, recent studies of biochemical genetic variation have found relatively large genetic differences between pollock sampled from extremes of the species range, but relatively small genetic differentiation within regions (Iwata 1973, 1975, 1977; Johnson 1977). Based upon

differences in allelic frequencies for the protein locus tetrazolium oxidase, Grant et al. (1978) concluded that the widely separated populations of the western and eastern Pacific are relatively genetically isolated and distinct. Only weak regional differentiation was observed, however, among pollock collected from north and south areas of the eastern Bering Sea and from the Gulf of Alaska.

Other investigators have proposed a more complex population structure in the Bering Sea. From analyses of morphometric characters, Serobaba (1977) recognized four principal Bering Sea pollock populations: southeastern, northern, western, and Aleutian Island. Maeda (1972) proposed two distinct eastern populations based upon apparent migrations to separate spawning areas northwest and southeast of the Pribilof Islands. After comparisons of the age structure of commercial catches in northwest and southeast regions from 1963 to 1970, however, Chang (1974) concluded that the pollock population exploited by fisheries in the eastern Bering Sea is a unit stock.

Description of habitat

In the eastern Bering Sea, pollock occur widely distributed over the continental shelf, but show

Figure 33-2. The apparent density distribution of walleye pollock during August-October 1975, determined by a research vessel survey (Pereyra et al. 1976).

Figure 33-3. The apparent density distribution of walleye pollock during April-June 1976, determined by a research vessel survey (Bakkala and Smith 1978).

highest densities along the shelf edge (Figs. 33-2 and 33-3). The overall distribution pattern varies seasonally and between years, dominated by movements between deep and shallow water. Over the continental shelf pollock show a semidemersal behavior, tending to form schools near the bottom during daytime, then dispersing up into the water column at night. Along the outer continental shelf and slope, dense shoals are formed that may exceed 20-50 km in length and have mean densities of up to 9 mt per hectare.

Figure 33-4. Apparent vertical distribution of pollock densities on the continental shelf during August-October 1975 (based upon the data shown in Fig. 33-2).

Figs. 33-4 and 33-5 show two examples of the apparent depth distribution of pollock on the eastern Bering Sea continental shelf, determined from research vessel surveys. During August-October 1975, pollock were relatively symmetrically distributed around a median depth of approximately 200 m. During April-June 1976, the vertical pattern was less symmetrical but similar, although apparent densities were lower among all depth intervals except 400-450 m.

The domain within which pollock occur demersally

Figure 33-5. Apparent vertical distribution of pollock densities on the continental shelf during April-June 1976 (based upon the data shown in Fig. 33-3).

Figure 33-6. Temperature and salinity characteristics of the environment of walleye pollock in the eastern Bering Sea (shaded area). Pollock T-S data replotted from Kihara and Uda (1969); North Pacific Ocean water mass data are from Sverdrup et al. (1946).

in the eastern Bering Sea, then, is a broad section of the outer continental shelf and slope primarily within the 100-300 m depth range. The bottom morphology of the continental shelf in most of this region is featureless and level, with sand and silt surface sediments (Sharma 1974). In deeper water, the continental margin is steep and cut by large submarine canyons that substantially affect the flow and mixing of water currents along the shelf edge (Kinder et al. 1975).

In comparison to temperature and salinity characteristics of water masses in other regions of the North Pacific Ocean, pollock occur within a relatively extreme seawater environment in the eastern Bering Sea (Fig. 33-6). These distinctive conditions—cold and low salinity—result from mixing of shelf and oceanic waters within apparently broad zones of interaction along and over the eastern continental shelf (Takenouti and Ohtani 1974, Coachman and Charnell 1979). Source waters, mixing characteristics, and associated hydrographic features must all be important in determining the composition of both pollock prey and predators.

Stock size

Seven well-documented estimates of the absolute bulk biomass of eastern Bering Sea pollock are summarized in Table 33-1. A confusing feature of

these results is the use of different, and sometimes unspecified, geographical boundaries among the different analyses. In addition, because all of the estimates are based upon selective sampling (namely, demersal trawl nets, and in the case of estimates based upon commercial data, targeted fisheries), the relationships between statistical populations included in the analyses and true field populations are unclear. Management policies for eastern Bering Sea trawl fisheries assume a present pollock stock size of approximately 5 million mt (NPFMC 1978).

Table 33-2 and Fig. 33-7 show the observed variability and trends for five indices of the relative annual abundance (biomass) of pollock in the eastern Bering Sea. During the sixteen years of observations 1963 to 1978, apparent pollock densities have varied approximately ± 50 percent of the long-term mean values. Rather than showing large random year-to-year differences, most variations in index values seem to have represented longer-term trends. Judging by the longest time series (Bakkala et al. 1979b), pollock are relatively less abundant now in the eastern Bering Sea region than during the period 1965 to 1970.

The research survey data shown (see Table 33-2) as NMFS Crab-Groundfish are results determined from demersal trawl surveys conducted by the U.S. National Marine Fisheries Service (NMFS). A central 159,100 km² core area (Fig. 33-8) has been surveyed

TABLE 33-1

Summary of estimates of absolute population size for eastern Bering Sea walleye pollock.

Source	Region and time period ¹	Method ²	Estimated population ($\times 10^6$ mt)
<i>A. Based upon research survey data</i>			
Pereyra et al. 1976	Eastern Bering Sea shelf, Unimak Pass to 61° N (August-October 1975)	1	2.426
Bakkala and Smith 1978	Eastern Bering Sea shelf, Unimak Pass to 59° N (April-June 1976)	1	0.679
Okada 1978; Nunnallee 1978	Aleutian Basin (June-July 1978)	1	0.840
<i>B. Based upon commercial fisheries data</i>			
Chang 1974	Eastern Bering Sea shelf, INPFC areas 1 and 2 (1969-1970)	1	2.3-2.6
Chang 1974	Eastern Bering Sea shelf, INPFC areas 1 and 2 (1970)	2	2.3-2.4
Low 1974	Eastern Bering Sea, primarily INPFC areas 1 and 2 (1964-1971)	1	3.45-5.83
<i>C. Based upon model estimates</i>			
Laevastu and Favorite 1977	Eastern Bering Sea shelf	3	8.235

¹ A description of INPFC (International North Pacific Fisheries Commission) statistical areas is given in Forrester et al. (1978).

² Estimation methods: 1 = "area swept" (Baranov 1918; Alverson and Pereyra 1969); 2 = "cohort analysis" (Pope 1972), 3 = "model fitting," based upon commercial fisheries data.

Figure 33-7. Annual variations in apparent relative abundance, based upon the indices of Pereyra et al. (1976), updated to 1978, and Bakkala et al. (1979b) summarized in Table 33-2.

using uniform methods during approximately June 1 to August 15 of each year (Pereyra et al. 1976: Section VIII).

Figure 33-8. NMFS index area (shaded) for annual eastern Bering Sea crab and demersal fish population assessment surveys. Dots indicate routine sampling locations.

TABLE 33-2

Summary of indices of abundance of walleye pollock in the eastern Bering Sea, 1963 to 1978.¹

Source	1963	1964	1965	1966	1967	1968	1969	1970	1971	1972	1973	1974	1975	1976	1977	1978	Mean	s.d.	c.v.
A. Research survey data																			
NMFS Crab-Groundfish surveys (Pereyra et al. 1976: section VIII updated to 1978), units = mean kg/ha trawled (nominal).	—	—	—	—	—	—	—	—	—	—	29.2	27.5	21.1	46.7	37.7	47.1	34.9	10.7	0.31
B. Commercial fisheries data																			
Japanese fisheries statistics (Chang 1974), units = mt/hr combined gear.	3.09	2.72	4.80	4.40	7.09	7.67	7.79	13.1	—	—	—	—	—	—	—	—	6.33	3.38	0.53
Japanese fisheries statistics (Bakkala et al. 1979b), units = mt/hr pair trawl.	—	9.5	18.3	23.6	21.3	23.8	31.5	18.7	14.2	14.2	8.6	10.4	9.3	9.4	8.6	9.4	15.4	7.15	0.46
Japanese fisheries statistics (Low 1974), units = 10 ⁶ mt.	—	3.77	4.02	3.46	5.83	3.81	4.92	5.41	4.92	—	—	—	—	—	—	—	4.52	0.87	0.19
Japanese fisheries statistics (Ikeda et al. 1977; Okada et al. 1979), units = % of 1976 pair trawl catch per unit of effort.	—	—	—	—	—	—	—	—	—	—	121	102	91	100	86	93	98.8	12.4	0.12

¹ Mean = arithmetic mean; s.d. = standard deviation of mean; c.v. = coefficient of variation (= ratio of standard deviation to mean).

TABLE 33-3

Indices of age-class abundance measured by NMFS Crab-Groundfish research vessel surveys within a central area of the eastern Bering Sea during June to mid-August, 1973-78
(units = 10^6 individuals per 159,100 km²).¹

Survey year	Number of trawls	Number of otoliths	Age-classes (yr)														
			1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1973	111	490	756.9	518.2	146.0	193.4	99.4	34.4	91.5	117.8	18.9	17.9	4.51	—	0.401	0.072	—
1974	111	954	2840.6	850.3	287.9	56.1	67.7	38.0	39.1	29.0	29.5	5.12	3.00	0.176	0.190	—	—
1975	111	766	758.4	402.4	614.2	108.0	24.6	27.5	14.7	14.2	7.98	5.99	0.52	0.594	0.009	—	—
1976	107	1990	729.0	500.4	479.7	1014.1	132.5	35.0	38.8	46.2	41.0	22.3	8.28	1.850	0.607	0.032	0.171
1977	112	944	2241.9	630.2	145.8	245.3	231.9	72.0	29.8	23.7	23.6	13.1	11.9	3.200	0.491	0.180	—
1978	116	1256	1170.9	400.4	806.9	507.0	139.5	92.3	29.1	24.2	29.1	19.0	6.26	5.100	—	—	—
Overall mean (1973-78)			1416.3	550.3	413.4	354.0	115.9	49.9	40.5	42.5	25.0	13.9	5.75	1.82	0.280	0.050	0.030
Standard deviation (1973-78)			906.7	169.9	267.7	359.4	71.0	26.0	26.5	38.4	11.2	7.11	4.02	2.01	0.260	0.070	0.070
Coefficient of variation (1973-78)			0.64	0.31	0.65	1.02	0.61	0.52	0.65	0.90	0.45	0.51	0.70	1.10	0.93	1.40	2.33

¹Based upon estimates of nominal sampling effort, uncorrected for differences in effective fishing power. Italicized values were apparently affected by in- or out-migrations.

Size and age composition

Rather than being a homogeneous pool, the eastern Bering Sea pollock population exhibits a size and age structure that reflects pulsed annual birth inputs, birth and survival rates, and a differentiation of distributional patterns based upon age. Pollock are generally considered to spawn during only one period each year. In the eastern Bering Sea, the spawning period has been reported to extend from the end of February through June, with peak activity in May (Serobaba 1968). This concentration of spawning within a four-month period, loosely synchronized with the timing of spring plankton production, provides for a large annual pulse of eggs and larvae.

Table 33-3 and Fig. 33-9 summarize the apparent age frequency distributions observed for pollock within the central region of the eastern Bering Sea (shown in Fig. 33-8), determined from NMFS demersal trawl surveys during the period 1973-1978. The data represent an overall range in body size (fork length) of approximately 6-90 cm. Ages were determined from readings of saccular otoliths.

Features of the observed age structure include: (1) a maximum age of 15 years; (2) a relatively high overall average mortality rate; (3) variable age-class abundance between years; and (4) large, dominant year-classes were apparently not maintained over successive years. Fig. 33-10 shows the overall pattern of survivorship, based upon the mean apparent abundance of each age group (i.e., column means) during the six years of observations. Assuming that

Figure 33-9. Annual variations in apparent age frequency distributions, based upon the data in Table 33-3.

Figure 33-10. Overall survivorship curve, based upon the data in Table 33-3.

rates of in-migration to, and out-migration from, the index area were equal, and that:

$$N_2 = N_1 e^{-Zt} \quad (1)$$

where

- N_1 = the number of individuals at any moment,
- N_2 = the number of individuals at some later time,
- Z = an instantaneous rate of total mortality,
- t = the intervening time interval, and
- e = the natural constant 2.71828...,

then the overall observed instantaneous mortality rate

(Z) was 0.72/yr. This rate corresponds to 51 percent of the pollock population dying each year. Rather than being constant, mortality was relatively high for ages 1, 4, 5, and 10 or more years (Fig. 33-11). Factors that may have contributed to apparently higher

Figure 33-11. Age-specific instantaneous mortality rates, based upon the data in Table 33-3.

mortality at ages four and five include the possibilities of increased selective removal by fisheries, increased migratory activities associated with ontogeny, and effects of reproductive stress. Senescence was presumably a major cause of higher mortality at ages 10 and older. Life table characteristics are summarized in Table 33-4.

Contrary to the assumption of a closed population, or that in- and out-migrations were in equilibrium, the indices of abundance observed in 1975 for all age-classes except the age-one population were inconsistently low (relative to cohort abundances observed in 1976), apparently indicating a distributional shift out of the index area during that survey period.

Another interesting result was the indication of density-dependent mechanisms possibly regulating age-class size (Fig. 33-12). Although the relative

year-to-year variation in abundance of one-year-old populations was slightly less than the overall average of age-classes 1-12 years, the apparent sizes of two-year-old populations were relatively constant. The lack of evidence for persistence of strong year-classes—such as the 1973 and 1976 year-classes that were unusually abundant at age one—was also unexpected, particularly since cohort dominance seems to be a feature of the dynamics of at least some simple animal populations with a large cannibalistic adult fraction (Fox 1975).

The NMFS Crab-Groundfish and other research vessel surveys of the eastern Bering Sea continental shelf have also found a geographical differentiation of age structure (Pereyra et al. 1976: Fig. IX-34; Bakkala and Smith 1978: Fig. 38). In summer, low-density populations in central and inner areas of the continental shelf—particularly northeast of the Pribilof Islands—have tended to be almost exclusively composed of one-year-old individuals. Along the outer continental shelf and slope, populations have been composed of a complex mix of age-classes 2 to 10 or more years.

Growth

Although gross characteristics of the growth of eastern Bering Sea pollock have been described (Yamaguchi and Takahashi 1972, Chang 1974, Pereyra et al. 1976, Bakkala and Smith 1978), no studies have yet attempted to analyze or mathematically model details of the growth curve(s) or the energetics of growth processes. The approach taken here will be to analyze original data to examine

Figure 33-12. Comparisons of relative year-to-year variability in age-class size, based upon the data in Table 33-3. The coefficient of variation is used as the ratio of standard deviation to mean.

TABLE 33-4

Life table characteristics for walleye pollock, based upon NMFS Crab-Groundfish research vessel surveys, 1973-1978.

Age ¹ (yrs) x	Mean apparent ² abundance ($\times 10^6$) n_x	Survival ³ l_x	Mortality ⁴ (yr^{-1}) m_x
0.2-1.2	(5100.)	1.0000	—
1.2-2.2	1416.3	0.2777	-0.95
2.2-3.2	550.3	0.1079	-0.29
3.2-4.2	413.4	0.0811	-0.15
4.2-5.2	354.0	0.0694	-1.12
5.2-6.2	115.9	0.0227	-0.84
6.2-7.2	49.9	0.0098	-0.21
7.2-8.2	40.5	0.0079	(+0.05)
8.2-9.2	42.5	0.0083	-0.53
9.2-10.2	25.0	0.0049	-0.59
10.2-11.2	13.9	0.0027	-0.88
11.2-12.2	5.75	0.0011	-1.15
12.2-13.2	1.82	0.0004	-1.87
13.2-14.2	0.28	0.00005	-1.73
14.2-15.2	0.05	0.00001	-0.51
15.2-16.2	0.03	0.000006	—

¹ Assuming a median birthday of April 1.

² Number of individuals reaching age x at time t . The value for age 0.2 yr was extrapolated from the n_x values of ages 1.2 to 15.2 yr, assuming constant mortality.

³ Probability that an individual survives to age x .

⁴ Age-specific instantaneous mortality rate, computed as $m_x = \ln(n_{x+1}) - \ln(n_x)$. The value in parentheses at age 7.2 yr may indicate sampling error or effects of migration.

fundamental characteristics of growth in length, growth in weight, and growth and decay of cohort weight.

Table 33-5 summarizes the mean lengths of individuals in each age-class represented in Table 33-3, determined from annual NMFS demersal trawl surveys of the index area (shown in Fig. 33-8). To examine growth in length, a generalized form of the von Bertalanffy (1938) growth model was fitted to the overall mean lengths (Table 33-5—column means) of ages 1 to 13 years. The form of the generalized growth model used was:

$$Y_x = y_{\infty} [1 - e^{-K(x-x_0)}]^d \quad (2)$$

where

- Y_x = the length or weight at age x ,
- y_{∞} = an asymptotic value of length or weight,
- K = a relative growth completion rate,

TABLE 33-5

Summary of mean lengths-at-age determined from NMFS Crab-Groundfish research vessel surveys conducted during June to mid-August, 1973-78
(units = cm fork length)

Survey year	Age-classes (yr)														
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1973	15.7	25.7	32.9	39.2	42.0	43.4	46.0	47.3	49.5	52.4	56.1	—	63.5	66.0	—
1974	15.5	25.2	34.9	38.5	44.3	44.3	47.7	49.6	51.0	53.4	51.7	57.6	57.8	—	—
1975	14.3	24.3	32.5	39.6	44.3	47.1	50.5	53.4	55.3	56.2	55.6	63.4	64.0	—	—
1976	11.6	23.9	30.6	37.1	41.5	44.2	46.4	48.2	48.6	51.0	55.1	57.8	58.5	66.1	72.0
1977	15.9	23.9	33.7	41.7	44.8	47.5	50.0	51.7	52.0	53.3	54.4	54.5	52.0	60.0	—
1978	13.7	25.2	33.7	37.7	44.0	46.7	51.6	53.3	53.1	53.0	55.6	53.0	—	—	—

x_0 = a hypothetical age of zero size, and
 d = a dimensionless exponent reflecting absolute growth rate.

The model was fitted to the data by a computer program using a Taylor-series method for least squares approximation (Draper and Smith 1966). The results are shown in Figure 33-13, and the parameter values of best fit describing growth in length (sexes combined) were:

$$\begin{aligned} y_{\infty} &= 64.81 \text{ cm} \\ K &= 0.132/\text{yr} \\ x_0 &= 0.78 \text{ yr} \\ d &= 0.520. \end{aligned}$$

In summary, the pattern of growth in length observed was fast growth between ages 1 and 5, relatively rapid decrease in annual length increments with increasing age, and an asymptotic body length of approximately 65 cm.

To examine characteristics of growth in weight, the mean weights-at-age were approximated by applying a power relationship to the length data in Table 33-5. The relationship used was (Pereyra et al. 1976):

$$W = 0.0075 L^{2.977} \quad (3)$$

where:

W = estimated body weight in grams, and
 L = body (fork) length in cm.

After converting the mean length of each age-class to a computed "mean weight," for all years, the

overall mean weights (i.e., column means) were determined, and the generalized growth model (Equation 2) was fitted to the results (see Fig. 33-13). The parameter values of best fit describing growth in estimated weight were:

$$\begin{aligned} y_{\infty} &= 7057 \text{ g} \\ K &= 0.016/\text{yr} \\ x_0 &= 1.18 \text{ yr} \\ d &= 0.944. \end{aligned}$$

Figure 33-13. Growth in length and weight, based upon the data in Table 33-5. Symbols indicate the overall mean at each age.

In contrast to growth in length, growth in body weight was nearly linear within the age range 1 to 13 years, with almost constant increments in weight of 114 g/yr.

Table 33-6 summarizes the overall mean body sizes at each age and, combining the observed survivorship from Table 33-4 with mean weights-at-age, illustrates the apparent growth and decay of cohort weights. According to this model, a pollock cohort attains maximum biomass at approximately four years of age.

Reproduction

Because seasonal spawning aggregations and roe production have been of particular interest to commercial fisheries, major features of the reproductive biology of pollock have been fairly well described (Gorbunova 1954, Zverkova 1969, Serobaba 1968, 1971, 1974, Maeda and Hirakawa 1977). The attainment of sexual maturity is related to body size. On the basis of a classification of macroscopic gonad conditions observed during a large demersal trawl survey of the eastern Bering Sea during April to June 1976, Bakkala and Smith (1978) found that the

Figure 33-14. Length-maturity relationships observed from pollock taken during the April-June 1976 demersal trawl survey (Bakkala and Smith 1978).

TABLE 33-6

Characteristics of the growth of walleye pollock, based upon NMFS Crab-Groundfish research vessel surveys, 1973-1978.

Age ¹ (yrs) x	Mean body ² length (cm) y _x	Mean body ³ weight (g) w _x	Relative age class ⁴ weight (g) l _x w _x
0.2-1.2	—	—	—
1.2-2.2	14.4	21.9	6.08
2.2-3.2	24.7	105.2	11.35
3.2-4.2	33.0	251.0	20.36
4.2-5.2	39.0	409.6	28.43
5.2-6.2	43.5	565.0	12.83
6.2-7.2	45.5	650.9	6.38
7.2-8.2	48.7	796.6	6.29
8.2-9.2	50.6	892.6	7.41
9.2-10.2	51.6	945.4	4.63
10.2-11.2	53.2	1034.0	2.79
11.2-12.2	54.8	1125.0	1.24
12.2-13.2	57.3	1298.0	0.52
13.2-14.2	59.2	1436.0	0.07
14.2-15.2	(64.0)	(1800.0)	(0.02)
15.2-16.2	(72.0)	(2537.0)	(0.02)

¹ Assuming a median birthday of April 1.

² Overall observed mean fork lengths. The lengths in parentheses were based upon limited observations.

³ Overall mean predicted body weights (wet tissue).

⁴ Computed as the product of "survival" times "mean body" weight.

relationship between "fork length" and "proportion of individuals sexually mature" was best described by a sigmoidal curve of the model:

$$P = e^{-be^{-cL}} \quad (4)$$

where P is the proportion of the apparent population mature at size L cm, e is the natural constant 2.71828..., and b and c are constants. The fitted model for male pollock was found to be (Fig. 33-14):

$$P = e^{-725.947e^{-.224L}} \quad (5)$$

and the length at which 50 percent of the individuals were mature was 31.0 cm. Female pollock matured at longer lengths than males, and the fitted model was:

$$P = e^{-867.088e^{-.209L}} \quad (6)$$

with the length at 50 percent maturity equal to 34.2 cm.

In the eastern Bering Sea, the sexually mature fraction of the pollock population undergoes a

seasonal cycle of gonad development and reproductive activity (Yamaguchi and Takahashi 1972). Despite a relatively extended spawning period of February through June, the majority of the breeding population appears to ripen and spawn during April to mid-May. Within this spawning period, it is not yet clear whether a mature female releases eggs in one short pulse, or as multiple releases over a period of days or weeks.

As for many other fish species, attempts to measure individual female fecundity (i.e., number of viable eggs released per year) have largely been premature to an adequate understanding of the biological processes of oocyte development, oocyte release, and the extent of resorption (Foucher and Beamish 1977). This has resulted in uncertainty over which sizes of oocytes to include in estimates of annual zygote production, uncertainty of spawning model (namely, single- or multiple-spawning), and failure to assess the number of oocytes remaining in the ovary after spawning (Shew 1978).

The ovaries of female pollock contain oocyte populations often composed of two or three distinct size-classes (Gorbunova 1954, Serobaba 1974, Shew 1978). Although details of the production and fates of these different size groups remain to be studied, it seems likely that the small classes (oocyte diameters ca. 60-550 μm) may represent a reserve fund (Foucher and Beamish 1977) from which only a fraction develop each season to be spawned. Assuming that only oocytes in the largest size class (ca. 600-1500 μm diameter) are actually released during spawning and viable, the relationship Shew (1978) found between body size and potential fecundity was:

$$F = 0.29 L^{3.462} \quad (7)$$

where:

F = estimated number of oocytes in the 600-1500 μm diameter size-class, and

L = body (fork) length in cm.

By applying Equations 6 and 7 to the survivorship and growth data in Tables 33-4 and 33-6, estimates could be made of individual and age-class fecundity (Table 33-7). Interesting features of the results are that age-classes four and five years apparently contribute most to the potential reproduction of the population, two- and three-year-olds contribute little because of their small sizes, and the cumulative contribution of rare old (six years and older) individuals is large.

The behavior of pollock during the spawning period (February to June) apparently includes migrations to, and aggregations along, the outer continental shelf and slope (Fig. 33-15). Although

Figure 33-15. Locations of individuals observed in spawning condition during April-June 1976. Dots show exact positions, shading indicates inferred range of spawning activities.

spawning probably occurs along the entire eastern Bering Sea shelf edge, the outer shelf region just west and northwest of Unimak Island has been repeatedly identified as a major reproductive site (Maeda 1972, Maeda and Hirakawa 1977, Serobaba 1968, 1974). Trawl fisheries target upon pre-spawning and spawning populations, obtaining high catch rates. Some fishing vessels harvest pollock only for roe, discarding their catches after stripping the ovaries from ripe females.

Schools of pre-spawning and spawning pollock apparently move high into the water column, forming dense midwater layers (Takakura 1954, Serobaba 1974). Mating presumably involves pairing, and broadcasting of the pelagic non-adhesive eggs (ca. 1.5 mm in diameter) with no parental care. Consistent with reported locations of principal spawning areas, pollock eggs and larvae have been observed in highest concentrations along the outer continental shelf between the Pribilof Islands and Alaska Peninsula (Serobaba 1968, Waldron and Vinter 1978).

Although some studies of the early life history of pollock in the eastern Bering Sea have presupposed a simple counterclockwise current transport through the central shelf region during the first one or two years of age, recent data suggest a different and more complex pattern of movement (Figs. 33-16 and 33-17). As opposed to the previous impression of predominantly counterclockwise water motion over the eastern Bering Sea shelf, the present concept of

TABLE 33-7

Reproductive characteristics of walleye pollock,
based upon NMFS Crab-Groundfish research vessel surveys, 1973-1978.

Age ^a (yrs) x	Pivotal length (cm) ^b y_s	Proportion of females mature ^c P_s	Individual fecundity ($\times 10^3$) ^d b_s	Potential age- class fecundity ^e $l_s P_s b_s$
0.2-1.2	—	—	—	—
1.2-2.2	13.7	0.	0.	0.
2.2-3.2	23.5	0.008	16.2	0.02
3.2-4.2	31.4	0.289	44.1	1.09
4.2-5.2	37.0	0.641	77.9	3.59
5.2-6.2	41.3	0.842	114.0	2.72
6.2-7.2	43.2	0.901	133.0	1.39
7.2-8.2	46.3	0.947	169.0	1.33
8.2-9.2	48.1	0.963	193.0	1.54
9.2-10.2	49.0	0.970	206.0	1.10
10.2-11.2	50.5	0.978	229.0	0.69
11.2-12.2	52.1	0.984	255.0	0.35
12.2-13.2	54.4	0.990	296.0	1.17
13.2-14.2	56.2	0.993	331.0	0.03
14.2-15.2	60.8	0.997	435.0	0.004
15.2-16.2	68.4	0.999	654.0	0.004

^a Assuming a median birthday of April 1.

^b The size at time of spawning ("April 1") was computed as 95% the y_x value.

^c Values for ages 1 to 4 from Bakkala and Smith (1978), Table 26.

^d Number of oocytes ($\geq 600 \mu\text{m}$ diameter) produced per year, using the relationship from Shew (1978).

^e Potential age-class fecundity = product of survival (at time of spawning), proportion mature, and individual fecundity; l_s values were computed using the age-specific mortality rates, assuming equal survival of males and females. Units = 10^3 oocytes.

Figure 33-16. Observations of 0-group (two- to four-month-old) pollock during June to mid-August, based upon NMFS research surveys during 1978 and 1979.

Figure 33-17. Overall apparent range of one-year-old pollock during June to mid-August, based upon NMFS research surveys, 1975-79.

Figure 33-18. Apparent food web based upon pollock in the eastern Bering Sea. Dotted lines indicate pollock ontogeny. Solid lines show important feeding relationships, with arrowheads and line weights indicating the direction and relative magnitude of carbon flows.

dominant long-term mean circulation is an extremely slow (1 cm/sec) drift to the northwest approximately parallel to the bathymetry (Coachman 1979).

If spawning occurs predominantly along the southeast outer shelf, then several months of northwest drift would carry larvae to the vicinity of the Pribilof Islands. In fact, during NMFS Crab-Groundfish trawl surveys conducted from June to mid-August, 0-group pollock (approximately two to four months old, and 35-80 mm fork length) have been observed over a large area of the northwestern outer shelf (Fig. 33-16). Highest concentrations of 0-group juveniles, showing as dense midwater or near-bottom layers, have been noted directly west of the Pribilof Islands. Inspections of the stomach contents of demersal fish species throughout the eastern continental shelf region have found 0-group pollock (lengths 50-80 mm) to occur as an abundant food item only in this same area.

By one year of age, pollock are broadly distributed over the entire central and outer continental shelf, completely overlapping and extending inshore beyond the adult range (Fig. 33-17). By two years of age, pollock remain more restricted to deep water, and with growth, begin to recruit to the trawl fisheries.

FOOD-WEB RELATIONSHIPS

One of the most striking features of the eastern Bering Sea food web is the extent to which pollock are represented in feeding relationships and overall energy exchange (Fig. 33-18). During different life history stages, pollock feed upon a broad spectrum of prey (Takahashi and Yamaguchi 1972, Mito 1974, Clarke 1978, Feder 1978, Smith et al. 1978, Bailey and Dunn 1979). In turn, pollock serve as a major food resource for a wide variety of primarily high trophic level predators (Mito 1974, Feder 1978, Hunt 1978, Lowry et al. 1978, Smith et al. 1978).

Food and feeding

The feeding characteristics of pollock are apparently largely a function of body size, regional location, and time of year. Body size sets certain morphological and behavioral limits—such as size of the feeding apparatus, extent of dentition, and swimming speeds—that affect the types of prey that can be captured and ingested. Body size is also strongly related to vertical migratory behavior within the water column and geographical location—factors that

also determine available prey. Although regional, seasonal, and yearly variations in prey densities must be important in determining food composition and feeding rates, these have not yet been well described in the eastern Bering Sea.

Larval pollock (5-20 mm) have been reported by Clarke (1978) to undergo shifts in selection of prey sizes during ontogeny. The gut contents of small larvae (5-10 mm) consisted mainly of copepod eggs and nauplii. Larger larvae showed progressively increasing fractions of cyclopoid copepods, calanoids, and larval euphausiids.

Juvenile pollock (2-20 cm) presumably feed on a corresponding series of primarily zooplankton prey of increasing sizes, perhaps foraged from areas high in the water column. Although not well studied, important food items probably include copepods, amphipods, juvenile and adult euphausiids, eggs, and larval fishes.

As adults (lengths 20-90 cm) pollock are generalized predators that appear to follow a "diurnally schooling-nocturnally active" behavior pattern, although freshly ingested food items can be found in the gut at any time of day. Large eyes are apparently adaptive for visual feeding, hunting at night, and foraging during low winter light intensities. The sleek, fusiform body indicates capabilities of fast swimming speeds. A large mouth and abundant, fine needle-like dentition provide good apparatus for seizing and ingesting large, active zooplankton (primarily *Thysanoessa* spp. and *Parathemisto* spp.) and fish (mainly juvenile pollock) prey (Fig. 33-19).

Figure 33-19. Changes in prey selection with size (from Takahashi and Yamaguchi 1972).

To place the role of pollock in the eastern Bering Sea food web in perspective, estimates were made of the food requirements for three overall average stock densities and related to primary production (Table 33-8). At the average pollock densities that can be inferred from research vessel

TABLE 33-8

Comparisons of estimated food intake rates of walleye pollock in the eastern Bering Sea to primary production.

Pollock density ¹ (mg wet wt/m ²)	Pollock density ² (mg C/m ²)	Food intake rate ³ (mg C/m ² /day)	Food intake rate as percentage of primary production rate ⁴
2,500	133	2	1
5,000	266	4	2
10,000	532	8	4

¹ Overall absolute density; the italicized value was the mean density observed by Pereyra et al. (1976), assuming 100% trawl efficiency.

² Carbon equivalents were computed assuming a dry wt. of 14% wet tissue wt., and carbon fraction of 38% of dry wt. (Parsons et al. 1977: Table 12).

³ Assuming a mean feeding rate of 1.5% of body weight per day (Daan 1973).

⁴ Assuming a mean primary production rate (water column) of 200 mg C/m²/day (McRoy et al. 1972).

surveys (50-100 kg/ha), the estimated daily food requirements represented approximately 2-4 percent of average photosynthetic production.

Predation

Pollock are fed upon by a wide variety of other species, and cannibalism (or "intraspecific predation") is also high. To a certain extent, the pattern of predation upon pollock shows regional variations reflecting the range overlaps and activity areas of the different predator populations. In deep water along the outer continental shelf, juvenile and adult pollock are a major food item of Pacific cod, Pacific halibut (*Hippoglossus stenolepis*), Greenland halibut (*Reinhardtius hippoglossoides*), arrowtooth flounder (*Atheresthes* spp.), and sablefish (*Anoplopoma fimbria*). In the vicinity of the Pribilof Islands where large colonies of seabirds and pinnipeds occur, predation upon larval and juvenile pollock from the near-surface layer by seabirds—the Red-faced Cormorant, *Phalacrocorax urile*; kittiwakes, *Rissa tridactyla*, *R. brevirostris*; murre, *Uria aalge*, *U. lomvia*; puffins, *Fratercula corniculata*, *Lunda cirrhata*; and Parakeet Auklet, *Cyclorhynchus psittacula* (Hunt et al., volume 2)—is significant, and larger pollock are actively dived and fed upon by the northern fur seal (*Callorhinus ursinus*) (Harry, volume 2).

Other widely ranging predators apparently feed upon pollock throughout their eastern Bering Sea distribution: cottids (particularly *Myoxocephalus* spp. and *Ulca bolini*), the Steller sea lion (*Eumetopias jubata*), other seals (*Phoca* spp.) and toothed whales (perhaps particularly *Phocoenoides dalli* and *Orcinus orca*).

Although cannibalism upon eggs and larvae during the post-reproductive period is a common behavior of many fish species, pollock are unusual due to the extent that intraspecific predation is a normal activity and represents a large fraction of total food intake. It is not unusual to find 35-cm pollock feeding upon 5 to 10-cm juveniles, and 60 to 80-cm adults feeding upon pollock 10-30 cm. This peculiar feeding strategy—by which cannibalistic adults gain a meal and eliminate potential competitors at the expense of their juveniles—may result in sensitive density-dependent processes for regulating both the age structure and total size of the population (Fox 1975).

FISHERY CHARACTERISTICS

Present status and history

Detailed reviews of the history and character-

istics of the different eastern Bering Sea trawl fisheries are given in Pruter (1973, 1976), Forrester et al. (1978), Low and Akada (1978), NPFMC (1978), and Bakkala et al. (1979a). In summary, large trawl fisheries for walleye pollock have been a relatively recent development (Table 33-9), representing a shift from targeting upon other fish species—primarily Pacific ocean perch (*Sebastes alutus*) and yellowfin sole (*Limanda aspera*)—prior to 1964. This shift was a consequence of declining catch rates (i.e., apparent overfishing) of the other species, and the development of new processing techniques and end products designed to improve the world market demand for pollock.

During the period 1966-70, annual pollock catches increased steeply (see Table 33-9). Peak yields were taken during 1971-73, then catches declined to the present levels of 950,000-980,000 mt/yr. These recent catch levels represent management quotas established under the authority of the U.S. Fishery Conservation and Management Act of 1976, Public Law 94-265. During 1978, 84 percent of the total eastern Bering Sea pollock catch was taken by Japanese, 9 percent by U.S.S.R., 6 percent by Republic of Korea, and 0.3 percent by Taiwanese vessels. The present fishing fleets include factoryships, factory stern trawlers, independent stern trawlers, pair trawlers, and refrigerator transports.

Selective characteristics

Contrary to general impressions, trawl fisheries in the eastern Bering Sea are composed of specialized units (i.e., national and company fleets, and individual fishing vessels) that selectively target upon specific fish populations (particular species and size mixes) so as to maximize the economic return from the particular catching, processing, and storage capabilities of each vessel. The market products that a fishing vessel is oriented towards—frozen products (fillets or blocks), minced fish ("surimi"), meal, or roe—largely determine fishing strategies. Depending upon seasonal variations in the availabilities of fish populations, different fishing units may: (1) exclusively target their trawling activities upon pollock; (2) switch the direction of trawling among several higher-valued species, taking pollock incidentally; or (3) at times, even discard mistakenly-caught pollock as a trash item.

Fig. 33-20 shows the overall geographical distribution of pollock catches (total annual yield) by Japanese fishing vessels in the eastern Bering Sea during 1977. In general, vessels targeting specifically upon pollock would probably have mainly fished along the

TABLE 33-9

Summary of annual removals of pollock from the eastern Bering Sea by trawl fisheries, 1964 to 1979 (metric tons).¹

Year	Japan	U.S.S.R.	Republic of Korea	China, Taiwan	Poland	Total
1964	174,792	0	0	0	0	174,792
1965	230,551	0	0	0	0	230,551
1966	261,678	0	0	0	0	261,678
1967	550,362	0	0	0	0	550,362
1968	700,981	0	1,200	0	0	702,181
1969	830,494	27,295	5,000	0	0	862,789
1970	1,231,145	20,420	5,000	0	0	1,256,555
1971	1,513,923	219,840	10,000	0	0	1,743,763
1972	1,651,438	213,896	9,200	0	0	1,874,534
1973	1,475,814	280,005	3,100	0	0	1,758,919
1974	1,252,777	309,613	26,000	0	0	1,588,390
1975	1,136,731	216,567	3,438	0	0	1,356,736
1976	913,279	179,212	85,331	0	0	1,177,822
1977	868,732	63,467	45,227	944	0	978,370
1978 ²	821,306	92,714	62,371	3,040	0	979,424
1979 ³	(774,630)	(60,000)	(85,000)	(5,000)	(25,000)	(950,000)

¹ Data from Bakkala et al. (1979b), unless otherwise cited.

² Preliminary estimates.

³ Foreign fishing allocations (Pileggi and Thompson 1979).

Figure 33-20. Geographical distribution of the total annual Japanese pollock catch during 1977, summarized by $\frac{1}{2}^\circ$ latitude and 1° longitude squares.

outer shelf. Vessels taking pollock incidentally to other species, such as yellowfin sole, Alaska plaice (*Pleuronectes quadrituberculatus*), and deep pleuronectids (*Reinhardtius hippoglossoides* and *Atheresthes* spp.), are represented by the extended distributions of low-level catches over deep water and in the central shelf.

Although pollock are taken in all months of the year, a majority of the annual catch is taken during the summer from June through September.

Fishing efforts directed toward pollock are highly size-selective (Chang 1974: Fig. 12), a consequence of the fish densities (more specifically, catch rates) available at different body sizes, trawl mesh sizes, and processing requirements of the different vessels. Fig. 33-21 shows the size frequency distributions observed from "survey" and "fishery" populations during 1978. Compared to the size distribution of the pollock population determined from the more evenly weighted and inshore survey coverage of the eastern shelf, trawl fisheries were selecting relatively larger individuals at the shelf edge.

A disturbing recent trend of eastern Bering Sea pollock fisheries, however, has been the selection

Figure 33-21. Comparison of the size frequency distributions of survey and fishery pollock populations. Survey results are from the 1978 NMFS Crab-Groundfish sampling survey of the index area shown in Fig. 33-8. Fishery results are from 1978 NMFS Foreign Fisheries Observer Program sampling of the commercial catches of Japanese mother-ships, Japanese large stern trawlers, and Korean large stern trawlers in INPFC area 1.

of an increasingly larger proportion of small and young fish. The size distribution of the 1978 fishery catch shown in Fig. 33-21 represents an age composition of approximately 2 percent one-year-olds, 20 percent two-year-olds, 40 percent three-year-olds, 18 percent four-year-olds, and 20 percent five or more years of age. In the perspective of the apparent growth and reproductive schedules for the population (see Tables 33-6 and 33-7), it would appear prudent to reduce fishing pressures upon age-classes 1 to 3 years.

SEASONAL AND INTERANNUAL VARIABILITY

Seasonal variations

The eastern Bering Sea environment is characterized by strong seasonality. Temperatures are seasonally cyclic and extremely cold during winter. Mixing and overturn of the water column occurs as a result of heat loss from surface layers during the winter, followed by increased water column stability during summer warming. Biological production processes follow a cycle of pulsed high activities during spring and summer, followed by a long winter period of low activity.

Apart from these generalizations, distinctive biological domains (i.e., geographical subregions

and water mass layers) are apparently formed in the eastern Bering Sea characterized by differences in mean and extreme values of environmental conditions, relative variability, and timing of events. These different domains, or habitats, are presumably used by biological species according to their resource value (in terms of Darwinian fitness) at specific times.

If we broadly define migration as "an act of moving from one spatial unit to another" (Baker 1978), then walleye pollock can be seen to exhibit a continuum of movement types. At short time scales (hours or days), the movements of individuals reflect the daily activity pattern—forming and dispersing from schools, searching for food, and avoiding predators. At longer time scales (weeks and months), individuals and large subpopulations (such as certain age-classes) show movements between different biological domains—from deep water up onto the continental shelf and back, along the shelf, and between major regions.

A dominant behavior of pollock appears to be seasonal movements on- and off-shelf (i.e., movements normal to bathymetry). During winter months, shelf populations appear to retreat to deep water along the outer shelf edge and may extend pelagically out over deep water. During the summer, vertical and onshore movements occur that result in high concentrations of adults along the outer shelf and a dispersal of primarily juveniles (age-classes one and two years) into the central and inner shelf regions. These seasonally-related bathymetric shifts are reflected in the trawling depths of the commercial fishery (Fig. 33-22).

Since pollock are ectotherms (body temperatures in equilibrium with their surroundings) on- and off-shelf migrations appear to be an adaptive response to the extremely cold temperatures (0.0 to -1.7°C) of the shelf domain during winter. Along the shelf edge at depths of 200-300 m, water temperatures are relatively constant— $3-5^{\circ}\text{C}$ throughout the year, providing a warm winter refuge (i.e., freezing avoidance) layer. Dispersal from this layer out onto the continental shelf during summer presumably maximizes the exploitation of different food resources by different size- and age-classes.

Along-shelf movements (i.e., parallel to bathymetry) have been found from tagging studies and inferred from month-to-month changes in the patterns of commercial fisheries catches (Serobaba, 1970, Maeda 1972, Takahashi and Yamaguchi 1972). These have been suggested to represent the directed, seasonally-related movements of subpopula-

tions between exclusive wintering, spawning, and summer feeding areas. Although the outer continental shelf area just west and northwest of Unimak Island does appear to be a major reproductive site, other resource areas (including winter refuge, other low-level spawning areas, and summer ranges) now appear to be broadly distributed, and there do not seem to be clear long-distance migration routes to particular seasonal activity centers.

Two large subsurface domains that appear fundamentally different are the outer continental shelf areas northwest and southeast of the Pribilof Islands. Temperature conditions are colder in the northern area, winter sea ice extends seaward of the continental shelf, and winter overturn penetrates the water column to depths below the shelf edge (Favorite et al. 1977). As a result, in this northern region the timing of spring and summer events—movements related to spawning, the spawning period, and dispersal onto the shelf—appear to follow the southeastern region by a lag of approximately 1 to 2 months. A characteristic progression of the commercial fisheries from south to north along the outer shelf during spring to mid-summer may represent targeting upon time-lagged spring aggregations along the shelf edge, following a northward migrating subpopulation, or perhaps partly result from restrictions due to ice and weather conditions.

Evidence has been presented for a progressive decrease in the size composition of pollock in the southeast area during the post-spawning period, presumably due to movements of large adults from the reproductive site (Serobaba 1970, Yamaguchi 1979). Whether these same, migrating, spent individuals account for increases in population size composition that occur in the northwest area during the same time period, will need to be verified by mark and recapture studies.

The responses of pollock to winter ice cover may be expected to be largely determined on the basis of thermal characteristics of the subsurface (particularly bottom water) environment. Since an arctic distribution is not shown, and there is no evidence of the development of biochemical mechanisms for adapting to freezing conditions, then it must be assumed that pollock are excluded from shallow regions of the continental shelf during winter ice cover. Along the outer continental shelf, the penetration of relatively warm (1-5 C) subsurface circulation under ice cover may still provide an adequate thermal refuge.

As suggested by genetic similarities and some tagging studies, interregional exchange is apparently an important process that may also vary seasonally.

Figure 33-22. Monthly variations in fishing depths for pollock, based upon 1978 NMFS Foreign Fisheries Observer Program data from Japanese and U.S.S.R. independent stern trawlers in INPFC area 1. Dots indicate the median fishing depth during each month, vertical bars indicate the monthly range. The shaded area, depths 135-245 m, represents the principal range of values during the year. Only gear operations in which pollock represented 25 percent or more of the total catch (by weight) were included in the analysis.

A relevant research problem is the question of determining the origins and relationships (to shelf populations) of pelagic walleye pollock found over deep water in the central Bering Sea. During an acoustic survey of the Aleutian Basin during June and July 1978, Okada (1978) observed "spotted type" echo patterns from large pollock throughout essentially the entire region, mainly from midwater layers at approximately 30-80 m depths during daytime. Pollock in these layers were extremely uniform in size, with 82 percent of all individuals sampled between 44 and 50 cm.

Variations between years

Even if the eastern Bering Sea were a more constant physical environment, important properties of the walleye pollock population (e.g., abundance, size and age structure, growth, food requirements) might still be expected to undergo long-term fluctuations resulting from (1) internal processes of population regulation, and (2) the external, and variable, influences of predators (including fisheries) and prey. Density-dependent processes of population control may be expressed as variations in life table characteristics. Because the population is composed of approximately 15 age-classes, and because inter-age class predation (and perhaps competition) is significant, interactions between age-classes may alone result in inherent variations in age-class abundances (and

hence, total population density) between years. Similar control mechanisms must also result from the size-selective activities of predator populations, producing varying patterns of age-specific mortalities.

Particular properties of the population are apparently sensitive to certain age-classes. The weight density, or biomass, of the population will largely be dependent upon the abundances of age-classes two to five (see Table 33-6). Potential fecundity will primarily be dependent upon the abundances of age-classes four to nine (see Table 33-7).

The actual high year-to-year variability of the eastern Bering Sea physical environment adds other components that contribute to, and perhaps accentuate, fluctuations in population properties. Potentially important variations in physical environment include changes in general climate (McLain and Favorite 1976), anomalously extensive ice cover and formation of cold bottom water during winter, and changes in the flow and characteristics of eastern Bering Sea source waters.

Climate is apparently important in determining the timing and extent of seasonal migrations of pollock both on- and off- shelf. During warm years, spring movements into shallower water have appeared to occur earlier, and the density distribution of the population has appeared to shift farther inshore than during cold years (Pereyra et al. 1976: Figs. VIII-11 and VIII-12). The direct importance of thermal conditions in determining distributional characteristics of the population may be overemphasized, however, and other factors that vary with temperature (e.g., particular food items) may be a more direct cause.

Hydrographic conditions and short-term climate (weeks) presumably have large influences upon annual reproductive success by contributing high variability to the survival of eggs, larvae, and early juveniles. The effects of variable early survival upon year-class abundances at subsequent ages, however, may be dampened or obscured by later density-dependent mortality processes.

FUTURE RESEARCH

Because of the importance of pollock in the eastern Bering Sea and other regions of the North Pacific, there are needs for substantially increased research activities studying their basic biology, population dynamics and ecological relationships. In order of priority:

1. There is a need to develop mathematical simulation models to examine the behavior of the population dynamics of pollock under different assumptions

of biological interactions and process rates. Single-species age-structured models are needed to examine detailed processes of internal regulation. Multispecies models are needed to study effects of interactions between species. Yield-oriented models are needed to enable more objective evaluation of potential effects of different harvesting levels by fisheries, and to test consequences of feedback and non-feedback management policies. Simulations of environmental randomness should be included in all three types of models.

2. We need to ensure the collection of more uniform population data, and in particular, to strive for developing longer observational time series based upon consistent procedures. This information is needed for use in mathematical models, and will provide the basis for scientific management of the fisheries. Fisheries surveys of pollock in the eastern Bering Sea would be more productive in the long run if the participating agencies better coordinated their research efforts, standardized sampling methods and gear, and agreed upon uniform statistical areas. There is also a need to improve the uniformity and quality of data reported from the commercial fisheries.

3. There is a need to study relationships between spawning population size, reproductive success, and subsequent age-class abundances and survival rates. Size- and age-specific mortality rates need to be evaluated as potentially important density-dependent mechanisms regulating population age structure and size.

4. The interrelationships of pollock with other species, particularly marine mammals, need to be studied.

5. We need a better understanding of seasonal movements and population exchange: (1) along the eastern Bering Sea continental shelf, (2) between populations associated with the eastern Bering Sea continental shelf and pelagic populations over deep water in the Aleutian Basin, and (3) between major geographical regions.

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Population Characteristics and Ecology of Yellowfin Sole

34

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ABSTRACT

Yellowfin sole is a major component of the demersal fish community of the eastern Bering Sea continental shelf; in 1975, it made up 64 percent of the total flounder biomass and 23 percent of the total sampled fish biomass. It has been fished commercially on a regular annual basis since 1954. Intense exploitation in 1959-62 (when 1.6 million mt were harvested) was the apparent cause of a severe decline in abundance; an estimated virgin exploitable biomass of 1.6-2.0 million mt fell to 0.8 million mt in 1963. The biomass remained at this lower level through the early 1970's, but since 1972 it has shown a marked increase and may have reached 1.4 million mt in 1978. This increase stems from the recruitment of a series of abundant year-classes originating in the years 1966-70.

Yellowfin sole migrate seasonally from outer continental shelf and slope waters (>100 m) occupied in winter and early spring to inner shelf waters (15-75 m), where spawning occurs in summer. Offshore migrations by adults in fall and winter are an apparent response to the ice cover and cold water temperatures that characterize inner and central shelf waters in winter.

Unlike the adults, the young remain in shallow nearshore nursery areas throughout their first few years of life. They begin to disperse to more offshore waters at three to five years of age, and by five to eight years of age they occupy much the same waters as older fish.

Yellowfin sole grow slowly and at eight to nine years of age reach a length of 25 cm, which is about the average size of fish taken in the commercial fishery. They are first recruited to the fishery at four to five years and are fully recruited at seven years, the age when the exploitable stock reaches its maximum biomass. Growth declines between ages seven and eight, when a high proportion of the population reaches sexual maturity. The rapid increase in mortality after the ninth year is apparently due to spawning stress.

Climatic variations in the eastern Bering Sea since the mid-1960's have produced some apparent changes in year-class abundance and in distributions and migrations of yellowfin sole. Abundant year-classes were produced in years of relatively warm climatic conditions; year-classes produced in cold years had relatively low abundance. Evidence also suggests that extensive ice cover may delay the start of spring inshore migrations, and residual cold water in central shelf regions may alter patterns of summer migrations and distributions.

INTRODUCTION

Yellowfin sole (*Limanda aspera*) is a right-eyed flounder of the family Pleuronectidae (Fig. 34-1), one of two species of *Limanda* found in the eastern Bering Sea, the other being the longhead dab (*L. proboscidea*). Yellowfin sole is a major component of the Bering Sea ichthyofauna; the longhead dab is a minor nearshore species. Yellowfin sole is a relatively small flounder; commercial catches in recent years reflect an average size of 25 cm and 175 g. Fish of 25 cm are eight or nine years old; however, maximum lengths of 53 cm (Ellson et al. 1950) and ages in excess of 20 years (Wakabayashi 1975) have been reported.

The yellowfin sole is limited in distribution to continental shelf and slope waters of the North Pacific Ocean, the Bering Sea and, to a limited extent, the Chukchi Sea (Fig. 34-2). It ranges along the Pacific Coast of North America from Barclay Sound (about 49°N), Vancouver Island, British Columbia, northward into the Chukchi Sea (Hart 1973), and along the Asian coast from the Gulf of Anadyr southward to the east and west coasts of Hokkaido Island, Japan, and along the Asian mainland in the Okhotsk Sea and the Sea of Japan to about 35°N off South Korea (Fadeev 1970a). Its bathymetric range is from about 5 to 360 m, although in some regions (e.g., the Gulf of Alaska) it is limited to continental shelf waters of generally 100 m or less. The deepest recorded occurrence (360 m) is in the eastern Bering Sea.

Yellowfin sole reach their maximum abundance in the eastern Bering Sea; they are by far the most abundant flounder in this region. An extensive trawl

Figure 34-1. Yellowfin sole (illustration from Hart 1973, courtesy Environment Canada, Ottawa).

survey in 1975 (Pereyra et al. 1976) found the biomass to be approximately 1.0 million mt, representing 64 percent of the total flounder biomass on the eastern Bering Sea continental shelf and 23 percent of the total sampled fish biomass. After walleye pollock (*Theragra chalcogramma*), it was the next most abundant demersal fish.

Although it is the predominant flounder in the eastern Bering Sea, yellowfin sole becomes a relatively minor species among the flounder complex in the Gulf of Alaska. It ranked eighth in relative abundance in flounder catches in the northern Gulf of Alaska (Hughes 1974) and is only occasionally taken in southeastern Alaska waters (Schaefers 1951, Ellison and Livingston 1952).

The abundance of yellowfin sole is lower in U.S.S.R. waters than in the eastern Bering Sea, but concentrations are large enough in certain regions to attract commercial fisheries (see Fig. 34-2).

Yellowfin sole and other groundfish of the eastern Bering Sea were first exploited commercially by Japan, initially by exploratory vessels in 1930 and then by a mothership-catcher boat operation off Bristol Bay in 1933-37 and 1940-41 (Forrester et al. 1978). In the 1933-37 period pollock and flounders were taken for reduction to fish meal; catches ranged

up to 43,000 mt annually. In the second period the fishery targeted on yellowfin sole for human consumption, with catches of all species ranging from 9,600 to 12,000 mt in the two years.

After World War II, Japanese distant water fisheries resumed operations in the eastern Bering Sea in 1954 with mothership and independent trawler fleets targeting on yellowfin sole. In the initial period of this fishery (1954-57), catches ranged from about 12,500 to 24,700 mt (Table 34-1).

In 1958 the U.S.S.R. also entered the fishery, and from 1959 to 1962 the exploitation of yellowfin sole was very intense with annual catches ranging from 185,000 to 554,000 mt. There is evidence that catches of this magnitude severely reduced the abundance of the stock (Wakabayashi et al. 1977). In the subsequent period of 1963-71, catches were much lower, ranging from 53,800 to 167,100 mt. Catches declined even further in the period 1972-77 as the U.S.S.R. discontinued its target fishery for yellowfin sole, presumably because of low catch rates. Catches in this period, primarily by Japan, ranged from 42,200 to 78,200 mt. In 1978 the U.S.S.R. resumed its target fishery for yellowfin sole; preliminary data indicate that all-nation catches rose to about 139,100 mt.

Figure 34-2. Overall distribution and areas of commercial fishing for yellowfin sole.

POPULATION CHARACTERISTICS

Stock biomass

The virgin biomass of exploitable yellowfin sole (age six and older) in the eastern Bering Sea has been estimated to range from 1.3 to 2.0 million mt (Alverson et al. 1964, Wakabayashi 1975). A virtual population analysis (Wakabayashi 1975) revealed that by 1963 the exploitable population was reduced to approximately 40 percent (801,500 mt) of the maximum estimate of virgin size. This decline occurred during the period of intense exploitation by Japan and the U.S.S.R., when 1.6 million mt were harvested in 1959-62. The analysis also showed some recovery of the resource in the mid-1960's, when the exploitable population was estimated to reach about 950,000 mt (Table 34-2). This was followed by another decline to about 790,000 mt in 1970.

A cohort analysis covering the period 1964-75 (Wakabayashi et al. 1977) showed trends in abundance similar to those indicated by the virtual population analysis, but yearly biomass estimates varied to some degree between the two studies (see Table 34-2). The cohort analysis showed the exploitable

population reaching a low of 604,000 mt in 1969 and then increasing to 910,000 mt in 1975. This increase in abundance has also been shown by indices of relative abundance from the commercial fisheries and from research vessel surveys (Bakkala et al. 1979). The estimated biomass in 1975 based on the cohort analysis was similar to that obtained from an extensive trawl survey of the eastern Bering Sea in the same year.

Indices of relative abundance and biomass estimates from trawl surveys have shown that abundance continued to increase through 1978 (Bakkala et al. 1979). The exploitable biomass may have reached 1.4 million mt by 1978, at least 70 percent of the estimated virgin population size. The primary reason for the increase has been the recruitment of a series of relatively strong year-classes originating in the years 1966-70 (Fig. 34-3). The cumulative contribution of these year-classes may have resulted in a doubling of the overall exploitable biomass over the period of 1973-78.

Stock structure

Yellowfin sole form dense concentrations on the outer continental shelf of the eastern Bering Sea in

TABLE 34-1

Annual catches of yellowfin sole in the eastern Bering Sea
(east of 180° and north of 54°N) in metric tons.

Year	Japan	USSR	ROK ¹	Total
1954	12,562	0	0	12,562
1955	14,690	0	0	14,690
1956	24,697	0	0	24,697
1957	24,145	0	0	24,145
1958	39,153	5,000	0	44,153
1959	123,121	62,200	0	185,321
1960	360,103	96,000	0	456,103
1961	399,542	154,200	0	553,742
1962	281,103	139,600	0	420,703
1963	20,504	65,306	0	85,810
1964	48,880	62,297	0	111,177
1965	26,039	27,771	0	53,810
1966	45,423	56,930	0	102,353
1967	60,429	101,799	0	162,228
1968	40,834	43,355	—	84,189
1969	81,449	85,685	—	167,134
1970	59,851	73,228	—	133,079
1971	82,179	78,220	—	160,399
1972	34,846	13,010	—	47,856
1973	75,724	2,516	—	78,240
1974	37,947	4,288	—	42,235
1975	59,715	4,975	—	64,690
1976	52,668	2,908	625	56,201
1977	58,139	284	55	58,478
1978	62,736	76,300	69	139,105

¹ 0 indicates no fishing, — indicates fishing but no reported catch.

winter. The largest of these is formed in the vicinity of Unimak Island and the second largest west of St. Paul Island; other lesser winter concentrations have also been recognized—one may be south or east of St. George Island and the other, consisting of small fish, in Bristol Bay (Fadeev 1970a, Wakabayashi 1974). Japanese tagging studies have shown that the wintering concentrations near St. George Island and Unimak Island combine in spring before onshore migration to areas off Bristol Bay (Wakabayashi et al. 1977). The wintering concentration of small fish in Bristol Bay is also thought to be a part of the Unimak Island/St. George Island group, perhaps representing a segment of the juvenile part of the population.

Figure 34-3. Biomass estimates for the exploitable population of yellowfin sole in the southeast Bering Sea. Year-classes for certain ages are shown with appropriate bars.

The second largest wintering concentration, located west of St. Paul Island, appears to remain relatively independent of the Unimak Island/St. George concentrations throughout the year. This group probably migrates inshore between St. Paul and St. Matthew islands and forms concentrations in the vicinity of Nunivak Island in summer.

The apparently independent movements and distributions of the St. Paul Island and Unimak Island/St. George Island populations have suggested the existence of independent spawning stocks of yellowfin sole in the eastern Bering Sea—a northern stock (St. Paul group) and a southern stock (Unimak Island/St. George Island/ Bristol Bay group) (Waka-

bayashi 1974, Wakabayashi et al. 1977). Japanese tagging studies through 1973 indicated only limited mixing of the two groups (Wakabayashi 1974) and some studies suggested differences in biological characteristics (such as growth rates, length-weight relationships, and egg diameters) between populations in the proposed north and south stock areas (Kashkina 1965, Wakabayashi 1974).

Other studies of biological characteristics, however, have not supported the two-stock concept (Fadeev 1970a, Wakabayashi 1974). In addition, returns from Japanese tagging studies since 1973 have shown greater intermixing of fish between the proposed stock areas (see Fig. 34-4 for results of Japanese tagging studies through 1977).

Results of biochemical genetic studies using electrophoretic techniques have provided no evidence of major genetic differences between fish from the proposed north and south stock areas; however, the samples on which the studies were based were not taken during the spawning season, when the two stocks (if they exist) would be most clearly segregated (Grant et al. 1978). In addition, the finding of similar gene frequencies between fish from the two areas is not positive evidence that differentiation has not occurred. For example, if the populations had become segregated in fairly recent times, genetic differences might not be easily detected with present electrophoretic methods.

Thus, although accumulating evidence from tagging and genetic studies tends to suggest the presence of a single spawning stock of yellowfin sole in the eastern Bering Sea, this question has not been entirely resolved.

Size composition

Yellowfin sole taken by U.S.S.R. research vessels in the eastern Bering Sea in the period 1958-64 ranged from 5 to 49 cm in length (Fadeev 1970a). The average size of fish taken by the commercial fishery was 26-27 cm in 1958 and 1959, years immediately preceding the period of intense exploitation. The dominant sizes according to Fadeev (1970a) were 24-30 cm. In recent years, the average size has been about 25 cm in the commercial fishery.

Length data taken on large-scale Northwest and Alaska Fisheries Center (NAFAC) trawl surveys in 1975, 1976, and 1978 (Fig. 34-5) show fish ranging from 5 to 44 cm and averaging 22.2-22.6 cm for combined sexes. More than 96 percent of the surveyed population were less than 30 cm long. A relatively high proportion of fish in the 12-16 cm range in 1978 represents the recruitment of the apparently strong year-class of 1973.

TABLE 34-2

Estimates of the exploitable biomass (age six and older) of yellowfin sole in the eastern Bering Sea in 10^6 metric tons.

Year	Virtual population analysis ¹	Cohort analysis ²	Trawl surveys ³
1959	2035.1		
1960	1924.0		
1961	1521.6		
1962	1054.6		
1963	801.5		
1964	856.2	912.5	
1965	872.8	960.7	
1966	948.1	969.0	
1967	946.5	879.0	
1968	862.7	635.4	
1969	880.8	604.0	
1970	786.7	720.8	
1971		648.2	
1972		660.0	
1973		849.1	
1974		761.4	
1975		910.2	991.9
1976			
1977			
1978			1414.6

¹Wakabayashi (1975)

²Wakabayashi et al. (1977)

³Pereyra et al. (1976); Bakkala et al. (1979)

Age composition

The observed age range of yellowfin sole since the early 1970's has been 2-19 years for males and 2-21 years for females (Pereyra et al. 1976, Bakkala and Smith 1978); however, about 97 percent of the sampled population were younger than 13.

Age data collected during NAFAC research vessel surveys and from the commercial fishery show that the population underwent marked changes in age structure during 1973-78 (Fig. 34-6). In 1973, a high proportion (86 percent) of the population sampled by research vessel surveys was made up of fish aged seven or younger. By 1978 these age groups represented only 36 percent of the sampled population, and fish of ages 8-11 represented 54 percent. These changes are the result of the recruitment and advancement through the population of the series of

Figure 34-4. Release and recovery locations of yellowfin sole tagged by Japanese research vessels in 1970 and 1971 and recovered by commercial fishing vessels (from Wakabayashi et al. 1977). Numbers of fish released are shown at each tagging location; a few fish (as indicated by light dashed lines) were released by commercial vessels. The heavy dashed line between St. George Island and Cape Avinof represents the line separating proposed north and south stock areas.

relatively strong year-classes originating in the years 1966-70. Research vessel data indicate that the 1971 and 1972 year-classes may not be as abundant as those of 1966-70 but the 1973 year-class may be of above average strength.

Size and age at maturity and recruitment to the fishery

Male yellowfin sole begin to mature at a length of 10.5 cm, females at 18.5 cm (Wakabayashi 1974; see also Fig. 34-7). The length at which 50 percent of the population is mature is 12.8 cm for males and 25.2 cm for females. All males are mature at 25 cm and most females at 30 cm.

On the basis of samples collected in 1959-64, Fadeev (1970a) reported that males first begin to mature at 12 cm and females at 16-18 cm but that 50 percent of the population reached maturity at 16-18 cm for males and 30-32 cm for females. Wakabayashi (1974) suggests that fish may have matured at a smaller size in 1973 because the population was less abundant than in 1959-64.

Females begin to mature at age six and 50 percent reach maturity at about age nine; all the females were mature at age 15 (Fig. 34-7).

Yellowfin sole first become recruited to the fishery at 13 or 14 cm, which corresponds to age four or five. They become fully recruited to the fishery at age

Figure 34-5. Length composition of yellowfin sole of the eastern Bering Sea as determined by research vessel surveys in 1975, 1976, and 1978.

Figure 34-6. Age composition of yellowfin sole as shown by data from NWAFC research vessel surveys in June-August and by U.S. observer samples from the Japanese flounder fishery in September-December. Year-classes for certain ages are shown above appropriate bars.

Figure 34-7. Maturity composition of yellowfin sole from the eastern Bering Sea in May-June 1973, by length for males, by length and age for females (after Wakabayashi 1974).

seven (Laevastu and Favorite 1978), about the age (seven or eight years) when the exploitable stock reaches its maximum size in weight (Wakabayashi 1975).

Length-weight relationships, growth, and mortality

Length-weight relationships obtained in 1975 and 1976 show that females are only slightly heavier than males (approximately 1-10 percent) from lengths of 20-30 cm (Table 34-3). Von Bertalanffy growth curves (Fig. 34-8) and their parameters (Table 34-4) illustrate the similarity of growth characteristics in males and females. Six-year means of observed lengths and calculated weights (Table 34-5) indicate

TABLE 34-3

Length-weight relationships for yellowfin sole of the eastern Bering Sea
(data of Pereyra et al. 1976, Bakkala and Smith 1978).

Sex	Year	Time Period	Sample Size	Length range (cm)	Length-weight coefficients		Predicted weight (g) at length		
							10 cm	20 cm	30 cm
Male	1975	Aug-Oct	701	7-40	0.0128	2.956	11.6	89.8	297.6
	1976	Apr-Jun	213	9-33	0.0208	2.779	12.4	85.6	264.2
Female	1975	Aug-Oct	844	8-40	0.0102	3.035	11.1	90.6	310.2
	1976	Apr-Jun	293	10-37	0.0168	2.870	12.4	91.1	291.9

TABLE 34-4

Parameters of the von Bertalanffy growth curve for yellowfin sole of the eastern Bering Sea
(data of Pereyra et al. 1976, Bakkala and Smith 1978).

Sex	Year	Number of age readings	Age range	Length range (cm)	Standard error of curve fit	Parameter		
						L_{∞}	K	t_0
Male	1975	609	0,8-15	0,17-30	1.50	40.79	0.11	0.22
	1976	507	0,4-12	0,13-33	0.31	31.88	0.17	-0.02
Female	1975	707	0,8-15	0,19-39	0.80	40.28	0.11	-0.09
	1976	600	0,4-14	0,13-36	0.66	32.23	0.18	0.10

that annual increments of length increase from age four to age six and then decline at older ages. Annual increments of weight also increase from age four to age six, but are relatively constant from seven to thirteen. Laevastu and Livingston (1978) have also shown a decline in the growth rate between ages seven and eight. They attribute this decline to the fact that a large portion of the population reaches sexual maturity at this age and energy is diverted from growth to development of sex products.

Mortality decreases rapidly with age in juvenile year-classes and reaches a minimum between ages five and nine, about the age of entry into the fishery (Laevastu and Livingston 1978). Mortality increases rapidly after the ninth year, when a high proportion of the population reaches sexual maturity. The higher mortality at ages 10 and above may be due to spawning stress (Laevastu and Livingston 1978).

Wakabayashi (1975), using the method of Alverson and Carney (1975), estimated instantaneous natural mortality (M) for fish age four and older as 0.25, corresponding to an annual mortality rate of 22 percent.

POPULATION ECOLOGY

Species interactions

An important predator of yellowfin sole in the eastern Bering Sea according to Novikov (1964) is the halibut (*Hippoglossus stenolepis*). He found a close relationship between the distribution of halibut and yellowfin sole during the summer and autumn and suggested that the movements of halibut are governed to a large degree by the movements of its principal prey, yellowfin sole. The incidence of yellowfin sole in halibut stomachs in his studies ranged from 33 to 70 percent and, in terms of composition by weight, from 30 to 55 percent over various areas of the southeastern Bering Sea. Although there must be other predators of yellowfin sole, particularly during larval and juvenile stages, they have not been documented.

Yellowfin sole are capable of feeding on a variety of animals, from strictly benthic forms such as clams and polychaete worms to zooplankton (mysids and euphausiids) to pelagic fish (capelin and smelt)

TABLE 34-5

Six-year means of observed lengths and calculated weights at age for yellowfin sole of the eastern Bering Sea from NWAFC survey data of 1973-78.

Age	Mean length (cm)	Annual increment (cm)	Mean weight (g)	Annual increment (g)
3	12.0		19.5	
4	14.1	2.1	32.3	12.8
5	16.6	2.5	51.7	19.4
6	19.3	2.7	81.5	29.8
7	21.0	1.7	103.9	22.4
8	22.4	1.4	126.4	22.5
9	23.9	1.5	154.2	27.8
10	25.2	1.3	181.0	26.8
11	26.4	1.2	206.3	25.3
12	27.4	1.0	230.3	24.0
13	28.7	1.3	265.2	34.9
14	29.1	0.4	278.3	13.1
15	29.4	0.3	286.6	8.3

Figure 34-8. Von Bertalanffy growth curves for yellowfin sole taken during spring 1976 (Bakkala and Smith 1978). Symbols indicate the mean length at each age.

(Pereyra et al. 1976). About 50 different taxa have been found in stomachs of yellowfin sole in the eastern Bering Sea (Skalkin 1963). The kinds of organisms consumed vary by season, area, and size of fish. Although feeding generally stops in winter, instances of fairly intense winter feeding have been recorded (Fadeev 1970a). During the onshore migrations in May and June 1971, 73 percent of the fish that had wintered near Unimak Island were feeding, but feeding intensity was lower for fish that had wintered near St. George Island (0.05 percent), St. Paul Island (19 percent), and in Bristol Bay (0 percent) (Wakabayashi 1974). They feed more intensely as they move onto the central shelf; diet varies by region, apparently depending on the availability of food organisms. Fadeev (1970a) suggests that they depend more on zooplankton when benthic organisms are scarce.

Contents of 2,357 stomachs taken over a broad area of the eastern Bering Sea (Table 34-6) show that the primary food items, representing 65 percent of stomach content by weight, were bivalves, amphipods, polychaete worms, and echiuroid worms. Polychaete worms and amphipods were the principal food items in smaller fish (10-20 cm), polychaete worms and bivalves and then echiuroid worms and amphipods in larger fish (20-30 cm), and bivalves and echiuroid worms in fish longer than 30 cm.

Recurrent group analyses (Fager 1957, 1963; Fager and Longhurst 1968) have been used to demonstrate species associations within the demersal community of the eastern Bering Sea (Kihara 1976, Mito 1977, Pereyra et al. 1976, Bakkala and Smith 1978). The procedure identifies species relationships on the basis of co-occurrence within samples. When joint occurrences are equal to or exceed 0.50, the species are considered to show affinity.

These four studies showed wide variation in the species that co-occurred with yellowfin sole. This variation might arise from differences in areas, time periods, and years of the surveys from which the data were analyzed. However, certain species co-occurred more consistently with yellowfin sole than others (Table 34-7): Alaska plaice were found to co-occur in all studies and rock sole and Pacific herring in more than one study. Pacific halibut was not one of the species showing close association with yellowfin sole based on recurrent group analysis as suggested in the food habits study of halibut by Novikov (1964).

Distribution and seasonal movements

Adults

Wakabayashi (1974) summarized Japanese commercial catch data from December 1967 to October

TABLE 34-6

Stomach contents (in grams) by size group of
yellowfin sole collected in the eastern Bering Sea
in 1970 (Wakabayashi 1974).

Food Item	Size Group			Total
	101-200 mm	201-300 mm	>300 mm	
Gadidae	—	60.3	26.8	87.1
Osmeridae	—	12.7	—	12.7
Ammodytidae	—	12.5	56.4	68.9
Other Pisces	—	36.4	18.4	54.8
Amphipoda	22.6	180.2	45.3	248.1
Euphausiacea	1.6	61.7	38.1	101.4
Macrura	3.7	22.8	24.1	50.6
Mysidacea	0.4	—	0.2	0.6
Brachyura	—	39.4	7.8	47.2
Anomura	1.7	51.5	18.3	71.5
Crangonidae	0.3	10.0	0.8	11.1
Polychaeta	31.4	360.9	63.1	455.4
Cephalopoda	—	—	1.7	1.7
Bivalvia	6.5	403.6	553.1	963.2
Gastropoda	—	1.3	10.4	11.7
Ophiuroidea	2.3	36.1	1.4	39.8
Scutellidae	6.4	33.2	17.2	56.8
Echiurida	9.2	245.4	398.4	653.0
Ascidia	0.2	13.7	4.2	18.1
Holothuroidea	—	34.8	3.5	38.3
Sand ¹	7.2	91.9	13.7	112.8
Others	12.8	108.2	163.6	284.6
Indistinct	12.9	64.2	27.7	104.8
Total	119.2	1,878.3	1,496.7	3,494.2
No. of Stomachs	275	1,708	374	2,357

¹ Possibly tubes of Polychaeta.

1968 to illustrate seasonal changes in fishing grounds. These data probably also generally illustrate seasonal changes in distribution of major concentrations (Fig. 34-9). In winter months (December-March), catches were concentrated near the 200 m isobath in the area west of the Pribilof Islands; but from the Pribilof Islands to Unimak Island, they occurred between 100-200 m. In May, major catches were made near the 100 m isobath or in even shallower areas. By June, the largest catches occurred between 50 and 100 m and in July and August near the 50 m

isobath. In September the main catches again shifted to deeper water (50-100 m) and in October the greatest proportion of the catches near Unimak Island were taken at the 200 m isobath.

More recent data from a large-scale NWAFC trawl survey in April-June 1976 illustrate the apparent routes of movement during spring onshore migrations of the two major wintering concentrations north of Unimak Island and west of St. Paul Island. The spring of 1976 was unusually cold (Bakkala and Smith 1978) and the ice edge in April 1976 approximated an extreme southern location based on observations of Potocsky (1975) over the seventeen-year period 1954-70. Because of the extensive ice cover, research vessel fishing operations in April were mainly limited to the outer shelf southeast of the Pribilof Islands. The Unimak Island wintering concentration was clearly evident north of Unimak Island immediately offshore of the ice edge (Fig. 34-10). Yellowfin sole were highly concentrated in this area with catch rates ranging up to 6,800 kg/km trawled. Bottom temperatures near the leading edge of this concentration were near 0 C; the densest portion of the concentration was in bottom temperatures of 0.1-2.0 C.

In May, as the ice edge receded, the Unimak Island wintering concentration shifted eastward towards Bristol Bay, apparently moving with the receding ice edge and 0 C isotherm (Fig. 34-10). In June (Fig. 34-11), a number of independent concentrations were observed east of the location of the principal concentration observed in May. The nearshore concentration in Bristol Bay was made up of small fish not encountered in offshore waters. Fadeev (1970a) believed that a concentration of juvenile fish he observed in this area in spring indicated overwintering in Bristol Bay.

Survey data in August to October 1975 (Fig. 34-11) and from earlier years (see Figs. VIII-23 to 29 of Pereyra et al. 1976), indicate that yellowfin sole form a large summer concentration in outer Bristol Bay between the 40 and 100 m depth contours. Tagging data suggest that these fish winter off Unimak Island and St. George Island (see Fig. 34-4).

The second major wintering concentration, located west of St. Paul Island, may have been under the ice in April 1976 (see Fig. 34-10). This assumption is based on its location in May, in an area covered by ice in April. In June, the concentration of fish between St. Paul and St. Matthew islands may represent a migration of this wintering group towards Nunivak Island; Japanese tagging data show that most of these fish are distributed in the vicinity of Nunivak Island in summer (see Fig. 34-4).

TABLE 34-7

Species showing close association with yellowfin sole as indicated by recurrent group analysis.

Authority	Season	Years of study	Species showing affinity with yellowfin sole
Kihara (1976)	Summer	1966-71, 1974	Alaska plaice (<i>Pleuronectes quadrituberculatus</i>) Rock sole (<i>Lepidopsetta bilineata</i>) Flathead sole (<i>Hippoglossoides elassodon</i>) Pacific cod (<i>Gadus macrocephalus</i>) Walleye pollock (<i>Theragra chalcogramma</i>) Cottidae Agonidae
Mito (1977)	Winter	1972, 1974-75	Alaska plaice (<i>P. quadrituberculatus</i>) Rock sole (<i>L. bilineata</i>) Yellow Irish Lord (<i>Hemilepidotus jordani</i>) Plain sculpin (<i>Myoxocephalus jaok</i>)
Pereyra et al. (1976)	Summer	1975	Alaska plaice (<i>P. quadrituberculatus</i>) Pacific herring (<i>Clupea pallasii</i>)
Bakkala and Smith (1978)	Spring	1976	Alaska plaice (<i>P. quadrituberculatus</i>) Pacific herring (<i>C. pallasii</i>) Sturgeon poacher (<i>Podothecus acipenserinus</i>) Capelin (<i>Mallotus villosus</i>)

Early life stages

Knowledge of the location and timing of spawning of yellowfin sole is based mainly on catches of eggs during plankton surveys; spawning adults have rarely been observed. Spawning begins in early July and probably ends in September (Musienko 1963). Eggs were observed over a broad area of the eastern Bering Sea shelf from off Bristol Bay to off Nunivak Island (Fig. 34-12), and densities indicated that spawning was most intense south and southeast of Nunivak Island. Depths of spawning ranged from 15 to 75 m.

Kashkina (1965) also observed spawning between Nunivak Island and St. Lawrence Island (see Fig. 34-12). My observations of females with "running" eggs in July 1979 indicate that spawning also occurs in northern Bristol Bay.

Eggs are pelagic, and laboratory studies of eggs from the Okhotsk Sea indicate that hatching occurs in about four days at 13 C (Pertseva-Ostroumova 1961). The lower threshold temperature for successful egg development was 4 C. Eggs have been encountered in the eastern Bering Sea at temperatures of 6.4-11.4 C (Kashkina 1965, Musienko 1963). At hatching, the prolarvae are small (2.2-3.1 mm) and transparent, have large yolk sacs, and are capable of swimming (Pertseva-Ostroumova 1961, Musienko

1963). About three days after hatching, the yolk sac is absorbed and the larvae begin to feed; at this stage they range in length from 3.3 to 3.8 mm.

The time between hatching and metamorphosis to the juvenile stage is unknown. Partially metamorphosed young have been found in plankton hauls at 16.5-17.4 mm in length; since larger juveniles have not been collected in plankton tows they are assumed to have begun a bottom existence in shallow near-shore waters.

Juvenile yellowfin sole of 5-10 cm (two- and three-year-olds) are first observed in research vessel bottom trawls in inshore regions (Fig. 34-13). These age groups have been observed in low abundance off Kuskokwim Bay, in Bristol Bay, and along the Alaska Peninsula. They subsequently disperse to the more offshore waters occupied by adults (Fig. 34-13), and at lengths of 16-20 cm (mainly five- to eight-year-olds) occupy much the same waters as the larger fish.

ENVIRONMENTAL INFLUENCES ON THE RESOURCE

The eastern Bering Sea continental shelf is subject to wide seasonal and annual fluctuations in environmental conditions. In winter and early spring much

Figure 34-9. Seasonal changes in Japanese fishing grounds for yellowfin sole, December 1967 to October 1968 (Wakabayashi 1974). Symbols represent the percentages of the total monthly catches (shown within the figure) taken in various $\frac{1}{2}^{\circ}$ latitude and 1° longitude statistical blocks.

Figure 34-10. Distribution and relative abundance of yellowfin sole in April and May as shown by a NWAFC trawl survey in 1976 (after Bakkala and Smith 1978). As indicated in the text, concentrations of yellowfin sole can be noted in the vicinity of Unimak Island, west of St. Paul Island, and east of St. George Island.

of the continental shelf may be covered by pack ice, but it is ice-free by late spring or early summer. Water temperatures under the pack ice are near the freezing point of seawater (-1.8°C) while bottom water temperatures on the shelf in summer may reach 10°C or more in shallow nearshore areas. Potocksky (1975) has documented the wide annual variation in distribution of pack ice in the eastern Bering Sea for the 17-year period 1954-70. Annual variations in surface and bottom water temperatures have also been observed (McLain and Favorite 1976, Maeda 1977, Bakkala and Smith 1978).

Climatic conditions (see Ingraham, this volume) in the eastern Bering Sea appear to vary by multiyear cycles rather than randomly (Kihara 1977, McLain and Favorite 1976, Maeda 1977). These cycles are governed by the direction of prevailing winds. In years of northerly prevailing winds, the distribution of pack ice extends farther south and Alaskan Stream water entering the southeastern Bering Sea from the North Pacific Ocean is prevented from intruding onto the continental shelf. Southerly prevailing winds, on the other hand, limit the southward advance of pack ice in winter and allow Alaskan Stream

Figure 34-11. Distribution and relative abundance of yellowfin sole in June 1976 and August-October 1975 as shown by NWAFC trawl surveys (after Bakkala and Smith 1978). As mentioned in the text, concentrations of yellowfin sole can be noted in Bristol Bay in June 1976 and on the inner shelf off Nunivak Island and in the southeastern Bering Sea in August-September 1975.

water to intrude onto the continental shelf in summer to replace residual cold water (0 C or less).

One multiyear temperature cycle in the eastern Bering Sea ended in 1976. Maeda (1977) has shown that temperatures increased from 1961 to 1967 and then decreased from 1968 to 1976. The 1971-76 period has been recognized as a particularly cold period in the eastern Bering Sea (McLain and Favorite 1976, Bakkala and Smith 1978). In 1977 and 1978, temperatures were again warmer than in the preceding six years.

Research vessel surveys to assess the condition of crab and groundfish resources in the eastern Bering Sea have been carried out by the NWAFC annually since 1971; a more comprehensive series of data is available from 1973. The 1973-78 series provides information on the distribution and abundance of yellowfin sole during the last four years of the cold phase of the recent climatic cycle and the initial two years (1977-78) of a warmer phase. Age data collected during the surveys also provides information on the strength of year-classes originating as far back

as the early 1960's. These data provide some evidence that the fluctuations in environmental conditions observed over the past several years may have influenced the distribution, migrations, and abundance of yellowfin sole.

Distribution and migration

Possible influences of pack ice on the distribution and migration of yellowfin sole are illustrated by data from an NWAFC survey in April-June 1976. Temperature conditions were unusually severe during the survey; bottom water temperatures in June of the survey period were the lowest recorded in the southeastern Bering Sea since 1966 (Bakkala and Smith 1978), and during April the ice edge approximated an extreme southern location based on 17 years of observations of pack ice distribution, 1954 to 1970 (Potocksky 1975). In April 1976, the leading edge of the large Unimak Island wintering concentration

Figure 34-12. Distribution of yellowfin sole eggs as shown by plankton surveys (A) in July 1962 (after Kashkina 1965) and (B, C) in July and August-early September 1958 (after Musienko 1963).

Figure 34-13. Distribution of yellowfin sole by size group, August-October 1975.

of yellowfin sole was near the ice edge, with extremely dense concentrations of fish immediately off the ice edge (Fig. 34-14). These data give the impression that the highly concentrated fish were waiting for favorable conditions to start moving onshore. As shown by later survey data (see Fig. 34-10), this concentration moved inshore in May, apparently following the receding ice edge. The apparent sequence of behavior in relation to the ice edge suggests that the Unimak Island wintering group avoided migrating under the ice and that ice cover in spring 1976 may have delayed inshore migrations.

The timing of spring inshore migrations is not well defined, although Fadeev (1970a) has observed them starting from late April to mid-May over the three-year period 1959-61. Ice-induced delays to spring migrations are probably infrequent and of relatively short duration. The ice must reach an extreme southern location in spring months, as it did in 1976, to interfere with migrations beginning in late April and early May.

In contrast to this apparent avoidance of ice cover by the Unimak Island concentration, other concentrations of yellowfin sole may at times inhabit waters

covered by pack ice. As shown earlier (see Fig. 34-10), the location of the St. Paul Island wintering concentration in May suggests that it was under the ice in April. That young yellowfin sole may winter under the pack ice in Bristol Bay has also been noted earlier.

Water temperatures may also influence seasonal movements and distributions. Offshore movements in fall and winter may be an avoidance response to the cold bottom temperatures existing over the eastern Bering Sea shelf in winter. Most of the population winters on the outer continental shelf and slope in temperatures of 3.5-6.0 C (Fadeev 1970b). Spring onshore migrations, however, are not restricted by relatively cold water on the shelf; they commonly occur from relatively warm waters (up to 3-4 C) of the outer continental shelf to colder waters of the central shelf (Fadeev 1970b). The non-avoidance of cold water is illustrated by the survey data in April 1976 (see Fig. 34-14). Although warmer water in the range of 2-3 C was accessible to fish in the Unimak Island wintering concentration farther offshore, they were mainly found in temperatures of 0-2 C.

Data from the series of NWAFC Crab-Groundfish surveys in June to mid-August 1973-78 illustrate the relationship of the distribution of yellowfin sole to temperature isotherms in early to mid-summer (Figs. 34-15 and 34-16). During this series of years, bottom water temperatures encompassing the surveyed distribution of yellowfin sole were variable, ranging from relatively low in 1974, 1975, and 1976 to relatively high in 1973, 1977, and 1978. Highest catch rates (>200 kg/10,000 m²) occurred in a rather wide range of temperatures from -1 C to 7 C. Fadeev (1970b) has also commented on the wide temperature range (0-10 or 11 C) inhabited by yellowfin sole in summer. His data showed highest catch rates in bottom water temperatures of 1-6 C.

Although high concentrations of yellowfin sole were observed in a wide range of water temperatures during the 1973-78 Crab-Groundfish surveys, there was also some indication of differences in distribution relative to temperature conditions. In summer 1978, a relatively warm year, the main concentrations occurred in northern Bristol Bay. In the coldest years of the data series, 1975 and 1976, a number of widely separated concentrations were observed, some far to the south and west of northern Bristol Bay. In 1973, 1974, and 1977, years of temperature conditions intermediate in the series, concentrations appeared to be located between the more northern waters occupied in 1978 and the more southern areas where some of the concentrations were found in the

Figure 34-14. Distribution of yellowfin sole in April 1976 relative to bottom temperature contours and the edge of the pack ice.

colder years of 1975 and 1976. These data imply that the rate of migration to more northern waters of the inner shelf may be slower in cold years than in warm years, or that summer distributions differ with temperature conditions.

Abundance

Accumulating evidence suggests that variations in water temperatures observed in the eastern Bering Sea may also affect the year-class strength of yellowfin sole. Maeda (1977) has shown predominant year-classes originating in years of relatively high temperatures in the eastern Bering Sea, and year-classes of below-average abundance originating in years of relatively low temperatures. More recent data appear to confirm this relationship. Bottom water temperatures taken in the southeastern Bering Sea in June of 1966-78 are used as a measure of temperature variation and to relate to year-class abundance (Fig. 34-17). Year-classes originating in the warmer years of 1966-70, when June bottom temperatures ranged

from 2.0 to 4.5 C, were relatively strong. In 1971 and 1972, when June bottom temperatures were colder (near 1.0 C), abundance of year-classes was much lower than in the previous five years. On the basis of this relationship, temperatures in succeeding years indicate that the 1973 year-class will be relatively strong, but that the 1974, 1975, and 1976 year-classes may be relatively weak. Preliminary data suggest that the 1973 year-class is, in fact, relatively strong; recruitment of the 1974-76 year-classes to research vessel gear is not yet adequate to measure their abundance.

Environmental conditions would be expected to have their greatest influence on survival of yellowfin sole during early life history stages. Yellowfin sole spawn from July to September; and although June temperatures would therefore not be expected to influence the survival of eggs, larvae, or juvenile fish directly, they may provide an index to the environmental conditions that will prevail during the subsequent spawning and early life history stages.

Figure 34-15. Distribution of yellowfin sole in relation to isotherms in June-August during the relatively cold years of 1974, 1975, and 1976.

Figure 34-16. Distribution of yellowfin sole in relation to isotherms in June-August during the relatively warm years of 1973, 1977, and 1978.

Figure 34-17. Mean bottom water temperature in June at 34 standard survey stations in the southeastern Bering Sea (International Pacific Halibut Commission, 1976a, b; 1977; 1978) and year-class strengths of yellowfin sole. Year-class strength is based on estimated recruitment of age-6 fish measured by NWAFC research vessel surveys.

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Trans-shelf Movements of Pacific Salmon

35

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ABSTRACT

Five species of Pacific salmon are produced in the rivers and streams tributary to the Bering Sea shelf. All spend a portion of their juvenile and adult lives as residents of the shelf. Although this residence is transitory, salmon compose a significant and highly variable portion of the total pelagic fish biomass of the shelf. Knowledge of the distribution and abundance of salmon on the shelf is vital to our future attempts to assess the impact of exploiting them upon other living components of the shelf and the impact of man's modifying the shelf environment on the salmon resource.

Both maturing salmon and juvenile salmon are present on the shelf from May through September, but their migration routes do not overlap appreciably. Juvenile salmon migrate seaward along the coast, eventually moving to offshore waters as their size increases. Maturing salmon remain in the offshore waters of the shelf until they are near their home river systems. The shelf distribution of maturing salmon appears similar for all species migrating to rivers in the same geographic areas. Chinook salmon are earliest to enter the shelf during both spawning and seaward migration; later come sockeye, chum, pink, and coho salmon, in that order.

Annual and seasonal variability in sea temperature and type and abundance of food appear to influence the distribution, growth, and, indirectly, the survival of salmon while they remain on the shelf. Significant gaps exist in our knowledge of the shelf distribution and dynamics of Pacific salmon. The greatest contribution to this knowledge can accrue from studies conducted between 168°W and the shelf edge from 56°N to 66°N.

INTRODUCTION

Five species of Pacific salmon, *Oncorhynchus* spp., are produced in the river systems tributary to the Bering Sea shelf. The sockeye salmon, *O. nerka*, is the most abundant species; next is chum salmon, *O. keta*; and then, in order, pink salmon, *O. gorbuscha*; chinook salmon, *O. tshawytscha*; and coho salmon, *O. kisutch*; (Table 35-1).

Salmon are anadromous, i.e., they mature in the ocean and spawn in fresh water. All salmon spend a

portion of their juvenile¹ and adult lives as residents of the Bering Sea shelf. Although they spend only part of their lives there, salmon compose a significant and highly variable portion of the total pelagic fish biomass of the Bering Sea shelf during spring and fall. Knowledge of the seasonal movements, migration routes, and magnitude of annual variations in the biomass of salmon while they are in this area is vital to future attempts to assess the impact of exploiting them upon other living components of the shelf ecosystem and of physically modifying the shelf environment upon the salmon resource.

BIOLOGY OF PACIFIC SALMON

All species of Pacific salmon have similar life histories but they differ in fecundity (Table 35-2), food habits, growth rate, migration patterns, fresh-water and ocean age, age and size at maturity, and time and location of spawning. From early summer to early fall, salmon return from the sea to spawn in the rivers and streams from which they originated. When they enter fresh water, they cease feeding and derive their nourishment from body stores. All Pacific salmon die after spawning.

Salmon eggs are deposited in gravel beds of rivers, streams, or lakes, and the eggs hatch during the winter. The young, known as alevins, remain in the gravel until their large yolk sacs have been absorbed and emerge from the gravel in the spring as fry. The greatest natural mortality occurs in fresh water during

¹ The term "juvenile" refers to seaward-migrating salmon that have entered estuarine or marine waters but have spent less than one year in this environment.

TABLE 35-1

Relative abundance (in thousands of fish) of Pacific salmon (*Oncorhynchus* spp.) produced in river systems tributary to the Bering Sea shelf as indicated by average of U.S. commercial catches 1961-77, inclusive^{a,b} and available Soviet catch statistics.^c

Area	Species					Total	Percentage of total of all areas
	Sockeye	Chum	Pink ^d	Chinook	Coho		
Bristol Bay	7,349.2	496.9	1,181.6	97.6	43.0	9,168.3	82.1
Gulf of Anadyr	150.0 ^e	410.5 ^f	9.9 ^g	scarce	scarce	570.4	5.1
Yukon River	0.002	315.6	0.54	98.1	14.2	428.4	3.8
Alaska Peninsula (north side)	285.3	75.2	8.8	4.7	32.8	406.8	3.6
Kuskokwim Bay	9.0	90.6	26.3	38.7	78.4	243.0	2.2
Norton Sound	0.02	114.5	59.5	3.0	5.8	182.8	1.6
Kotzebue Sound ^a	0.006	168.1	0.004	0.003	—	168.1	1.5
Arctic coast ^h	i	j	j	i	—	—	—
TOTAL	7,793.5	1,671.4	1,286.6	242.1	174.2		
% by species	69.8	14.9	11.5	2.2	1.6		

^aNo reported catches for Kotzebue Sound in 1961.

^bSource Alaska Department of Fish and Game, Juneau, Alaska.

^cPravdin (1940) and data available at TINRO.

^dEven years only for U.S. waters.

^eEstimated possible annual catch (Pravdin 1940).

^fAverage 1974-78 inclusive.

^gAverage 1975, 1977, and 1978.

^hNo estimates of number available.

ⁱUnreliable reports of this species in American arctic rivers (Walters 1955).

^jDocumented reports of this species in Siberian and American rivers (Walters 1955).

the early life stages, and is greatly influenced by environment.

The fry of some species of salmon proceed immediately to sea; fry of other species reside in fresh water for a few weeks or for one or more years. They migrate seaward across the shelf to the open ocean regions of the central and western Bering Sea and North Pacific Ocean, where they spend most of their marine life. Depending upon the species, salmon usually spend from a few months to several years at sea (see Table 35-2).

Growth is slow during freshwater life, very rapid during the first summer at sea, and slower thereafter. The growth rate of salmon in both fresh water and the ocean varies among species, brood years, and stocks of the same species. Regardless of the amount of time spent in fresh water, however, salmon attain most of their growth in the ocean.

ABUNDANCE OF PACIFIC SALMON ON THE SHELF

According to commercial fishery catch statistics, the rivers and streams flowing into Bristol Bay (see Table 35-1) at the southeastern terminus of the Bering Sea shelf produce the greatest biomass of salmon (Fig. 35-1); information from the early 1900's (Pravdin 1940) and recent data available at TINRO indicate that those flowing into the Gulf of Anadyr in the northwestern shelf area are a distant second. Pacific salmon are produced in certain rivers of the Siberian and American arctic coasts (Walters 1955), but estimates of their relative abundance are not available. Apparently chum salmon, *Oncorhynchus keta*, are sufficiently abundant in the lower Lena River, which enters the Siberian arctic, to support a commercial fishery. Salmon produced in arctic rivers must traverse the Bering Strait and the Bering Sea shelf.

TABLE 35-2

Life history of the five species of Pacific salmon in Alaska—exceptions are frequent
(adapted in part from Bailey 1969, Merrell 1970, and Hartman 1971).

Species of salmon	Freshwater habitat	Time spent in fresh water after emergence from gravel	Estimated average juvenile weight at time of seaward migration (g)	Time spent at sea (yr)	Age at spawning (yr)	Average adult weight ^f (kg)	Average number eggs per female (thousands)
Sockeye	short streams and lakes	12-36 months	8.7 ^a	1-4	3-6	2.27	3.5
Chum	short and long streams	1 month	0.37 ^b	2-4	3-5	2.96	3.0
Pink	short streams	1 day, usually	0.26 ^c	1	2	2.00 ^g	2.0
Chinook	large rivers	3-12+ months	4.3 ^d	1-6	3-6	7.26	4.0
Coho	short streams and lakes	12-14 months	29.0 ^e	1-3	3-4	3.04	3.5

^a Average for age 1.0 and 2.0 Bristol Bay sockeye salmon (National Marine Fisheries Service Auke Bay Laboratory 1979, unpublished data).

^b Hooknose Creek, British Columbia (from Table 5 of Bams 1970).

^c Southeastern Alaska pink salmon (Bailey and Pella 1976).

^d Yukon River (Salcha River) chinook salmon (Trasky 1974).

^e Karluk Lake (Kodiak Island) coho salmon (Drucker 1972).

^f Mean weight for Alaska Peninsula (north side), Bristol Bay, and Yukon River stocks combined. Source: INPFC Statistical Yearbook (1961).

^g Mean for Alaska Peninsula (north side) and Bristol Bay only.

Estimates of the annual abundance of maturing sockeye, chum, pink, chinook, and coho salmon returning to Bristol Bay between 1951 and 1979, inclusive, have been made by Rogers (1977 and personal communication). Estimates of the average total abundance of each species returning to rivers and streams in Bristol Bay and along the northern side of the Alaska Peninsula have also been made by Stern et al. (1976). No similar estimates have been made for maturing salmon returning to other western Alaska rivers and streams entering the shelf north of Bristol Bay, nor are estimates available for the total abundance of each species of salmon returning to rivers and streams entering the Gulf of Anadyr or for rivers and streams along the Siberian and North American arctic coasts. If the catch statistics in Table 35-1 indicate the relative contribution of each area to the total population of maturing salmon entering the shelf each year, the areas north and west of Bristol Bay contribute 14.2 percent of this population. The reliability of this figure, however, can be determined only through improved estimates of the commercial catches, subsistence catches, and escape-ment of maturing salmon in these areas.

On the basis of very limited data and several

Fig. 35-1. Index map showing place-names and locations of salmon-producing regions.

assumptions listed below, we can estimate the possible range in the total number and biomass (weight) of maturing salmon entering the shelf each year to be

between 5.6 and 65.9 million fish or 13,776-162,114 mt. The assumptions in making this estimate are as follows:

1. The catch statistics given in Table 35-1 indicate the contribution of each area of the shelf to the total shelf population of maturing salmon.
2. Annual fluctuations in the abundance of salmon produced in all rivers and streams entering the shelf follow annual fluctuations in abundance for Bristol Bay between 1951 and 1979, inclusive (Rogers 1977 and personal communication).
3. The weighted mean weight of mature salmon of all species combined is 2.46 kg per fish. This weight was calculated by summing the products of the mean weight for each species (see Table 35-2) multiplied by its contribution to the total shelf population of salmon (see Table 35-1).

The estimate is derived by dividing the low and peak estimates (4.6 and 54 million fish) of the numbers of maturing salmon returning to Bristol Bay between 1951 and 1979, inclusive, by the mean proportion of salmon (0.82) Bristol Bay contributes to the total shelf population of salmon each year (see Table 35-1).

Estimates have also been made of the numbers of juvenile sockeye, chum, pink, chinook, and coho salmon entering Bristol Bay each year between 1950-1976, inclusive (Rogers 1977 and personal communication). In addition, the average and peak estimates of all species of juvenile salmon entering the shelf from Bristol Bay and from streams along the northern side of the Alaska Peninsula for 1955 through 1975 have been made by Stern et al. (1976). Estimates of juvenile salmon entering the shelf from areas north and west of Bristol Bay have not been made. The low and peak estimates of juvenile salmon migrating seaward from Bristol Bay for 1955-76 numbered 119 and 770 million fish, respectively. If the same rationale used to calculate the abundance of maturing salmon is used for juvenile salmon, the annual probable low and peak numbers of juvenile salmon entering the shelf are 145 and 939 million, respectively. If 6.72 g is the weighted mean weight for all species of juvenile salmon at the time they enter the shelf, the estimated total weight of juvenile salmon on the shelf in a given year is between 974 and 6,310 mt. The weighted mean weight for all species of juvenile salmon was calculated by summing the products of the mean weight of each species (see Table 35-2) multiplied by its contribution to the total shelf

population of salmon (as indicated by the adult catch in Table 35-1).

SEASONAL MOVEMENTS OF SALMON ON THE SHELF

The marine life of Pacific salmon may be divided into three phases: ocean life, spawning migration, and seaward migration. The seasonal timing and length of each phase is species and stock specific and fluctuates, within limits, because of environmental variation.

Ocean life

When salmon are distributed throughout the vast area of the Bering Sea and North Pacific Ocean (French et al. 1975), they undergo rapid growth and attain sexual maturity. This period of ocean life may last from a few months to several years depending on the species of salmon (see Table 35-2) and ocean growing conditions. More than one generation of sockeye, chum, chinook, and coho salmon are present on the shelf during ocean life. Immature and maturing salmon of different races and ocean ages and from different continents or geographic areas may, depending on the season, become segregated from or intermix with one another (Royce et al. 1968). Immature chinook salmon (age 0.1)² apparently are abundant in the central Bering Sea near the edge of the continental shelf in June and July (Major et al. 1978). Immature chum salmon, older than age 0.1, have been captured in research gillnets in July in the western Bering Sea shelf and in the slope area near Cape Navarin and the Gulf of Anadyr (Nishiyama et al. 1968). A few immature chum and sockeye salmon have also been taken during research fishing on the Bering Sea shelf as far north as 62°N in late July (Yonemori 1967).

Research fishing and salmon tagging have been conducted on the Bering Sea shelf only during the spring, summer, and early fall—periods when juvenile salmon are migrating seaward and maturing salmon are migrating to spawning grounds. The species composition and proportion of immature salmon of age 0.1 or older, of maturing salmon, and of seaward-

²Number of scale annuli acquired in fresh water is indicated by the figure preceding the decimal, and the number of annuli acquired in the ocean is indicated by the figure following the decimal. Thus, an 0.3 fish is one whose scales reflect 3 annuli at sea (freshwater age unspecified), a 2.0 fish is one whose scales show 2 annuli in fresh water (ocean age unspecified), and an age 2.3 fish is one with 2 annuli in fresh water and 3 at sea. Total age (year of life) is obtained by adding one to the sum of the freshwater and ocean annuli. For example, an age 2.3 fish is six years old.

migrating juvenile salmon residing in the shelf area between November and April are unknown but the numbers are probably negligible. Young chinook salmon (judged by their 26-35 cm fork lengths to have been in their first year in the ocean) have been taken by a Japanese trawler fishing in mid-winter for walleye pollock, *Theragra chalcogramma*, in the eastern Bering Sea at the shelf edge near 56°N, 168°W and 56°N, 173°W (Major et al. 1978).

General spawning migration

The time of the spawning migration, from departure of sexually maturing fish from the high seas of the North Pacific Ocean and western and central Bering Sea until arrival at the mouths of their respective home streams or river systems, is species and stock specific. Maturing salmon are most abundant on the Bering Sea shelf from mid-May to early September (Table 35-3). Maturing chinook salmon enter the Bering Sea shelf earliest; later in order come sockeye, summer chum, pink, fall chum, and coho salmon. The length of time a given salmon stock is present and its distribution in the shelf area during spawning migration depend upon the geographic location of its home river system, the size and/or ocean age of fish composing the population, and environmental conditions that influence the rate and direction of migration.

Sockeye, chum, and pink salmon have been captured in varying numbers during U.S. and Japanese research fishing and tagging studies at many places throughout the shelf during spawning migration. Fewer chinook and coho salmon, however, have been caught than other salmonid species because chinook and coho salmon are less abundant and research fishing did not take place when they were at their peak. Research fishing and tagging experiments have taken place mainly between mid-June and late July, when sockeye and chum salmon are most abundant on the shelf. Chinook salmon begin entering the shelf area in mid- to late May, coho in mid- to late July. As a result, we have less information about the distribution and direction of movement of chinook and coho than of sockeye, chum, and pink salmon.

The distribution and direction of migration and relative abundance of all five species of Pacific salmon on the Bering Sea shelf are depicted in Figs. 35-2 to 35-6. These figures were prepared by plotting the locations of capture of each species on a chart of the Bering Sea shelf for all years for which data are available. For sockeye, chum, and pink salmon, certain areas of the shelf consistently yielded larger catches or larger catch-per-unit-effort values than

Fig. 35-2. Distribution of sockeye salmon during spawning migration, mid-June to late July.

Fig. 35-3. Distribution of chum salmon during spawning migration, mid-June to early August.

other areas. For chinook and coho salmon, similar data were available only for Bristol Bay, where these species have been captured most frequently. Direction of migration of each species was derived from the published results of tagging experiments and direction-of-movement studies. The probable direction of migration was based on the geographic location or proximity of the home river system of the species and the verified direction of movement of

TABLE 35-3

Inclusive dates of peak abundance of maturing Pacific salmon on the Bering Sea shelf during spawning migration (from commercial fishery and test fishing catches and migration rates).

	Sockeye	Chum (summer)	(fall)	Pink	Chinook	Coho
Estimated at shelf edge ^a	15 Jun-25 Jun	22 Jun-13 Jul	—	5 Jul-15 Jul	30 May-19 Jun	20 Jul-2 Aug
Estuaries or river entrances:						
Alaska Peninsula ^b	20 Jun-25 Aug ^c	4 Jul-25 Jul	20 Jul-20 Aug	1 Aug-15 Aug	1 Jun-20 Jun	10 Aug-10 Sep
Bristol Bay ^d	30 Jun-10 Jul	25 Jun-15 Jul	g	20 Jul-30 Jul	10 Jun-30 Jun	1 Aug-20 Aug
Kuskokwim Bay ^e	17 Jun-5 Jul	g	g	g	10 Jun-30 Jun	25 Jul-15 Aug
Yukon River ^c	g	11 Jun-9 Jul	24 Jul-10 Aug	15 Jul-10 Aug	13 Jun-28 Jun	9 Aug-29 Aug
Norton Sound ^f	g	30 Jun-16 Jul	g	3 Jul-25 Jul	10 Jun-30 Jun	3 Aug-25 Aug
Kotzebue Sound ^f	g	g	1 Aug-20 Aug	g	g	g

^aEstimated dates are for Bristol Bay salmon only. Dates are based on migration rates of 45 km/day for sockeye and pink salmon, 56 km/day for chum and coho salmon, and 60 km/day for chinook salmon during the last 30 days of spawning migration. A travel distance of 680 nautical miles from shelf edge to the head of Bristol Bay (the mouth of the Kvichak River) was used in these calculations. The 2,000 m isobath in the vicinity of the heaviest concentration of a salmon species was delineated as the "shelf edge."

^bA. R. Shaul, Alaska Department of Fish and Game, Cold Bay, Alaska, personal communication, 1979. Dates are based on commercial fishery catches.

^cThere are two types of sockeye salmon spawning migrations to river systems of the northern Alaska Peninsula, i.e., those that reach peak abundance in the fishery in a short period of time between 25 June and 10 July, and those that occur over a long period of time between 20 June and 25 August with no definable peak period of abundance. A. R. Shaul, Alaska Department of Fish and Game, personal communication, 1979.

^dC. P. Meacham, Alaska Department of Fish and Game, Anchorage, Alaska, personal communication, 1979. Dates are based on commercial fishery catches.

^eW. Arvey, Alaska Department of Fish and Game, Anchorage, Alaska, personal communication, 1979.

^fW. Arvey, Alaska Department of Fish and Game, Anchorage, Alaska, personal communication, 1979. Dates are based on commercial fishery catches.

^gLow abundance precludes estimates.

Fig. 35-4. Distribution of pink salmon during spawning migration, mid-June to mid-August.

Fig. 35-5. Distribution of chinook salmon during spawning migration, early June to mid-July.

other salmon species captured in the same area. For example, maturing sockeye and chum salmon appear to migrate eastward north, and maturing chum and pink salmon south, of St. Lawrence Island (Yonemori 1967) (see Figs. 35-3 and 35-4). Major spawning

streams of chum and pink salmon are northwest of St. Lawrence Island in Kotzebue Sound and east of St. Lawrence Island in Norton Sound. Spawning streams of chinook and coho salmon are also to the east and northwest of St. Lawrence Island.

Fig. 35-6. Distribution of coho salmon during spawning migration, early July to early September.

Spawning migration, sockeye salmon

Sockeye salmon are most abundant during the spawning migration in the southeastern shelf area (see Fig. 35-2). They are concentrated in two bands offshore, north and south of the Pribilof Islands, one bound for rivers on the north side of the Alaska Peninsula and the other for rivers flowing into Bristol and Kuskokwim Bays. As they migrate across the shelf toward the head of Bristol Bay, they progressively segregate according to their rivers of origin (Straty 1975). This segregation begins when the fish are still as far as 200 km from the mouths of their home river systems. Bristol Bay sockeye salmon remain offshore, rather than entering coastal waters, until they are within 32-80 km of the mouths of their home river systems. Sockeye salmon bound for rivers on the north side of Bristol Bay (i.e., those rivers entering Nushagak and Togiak bays) are apparently more abundant in the northern than in the southern portion of the eastern shelf distribution.

Sockeye salmon captured north of St. Lawrence Island seem to be migrating eastward (Yonemori 1967). These fish are probably bound for the few sockeye salmon producing river systems north of St. Lawrence Island in Norton and Kotzebue Sounds (see Fig. 35-2). Northward migration of sockeye salmon east of St. Lawrence Island is uncertain because only one has been caught in this area (Hokkaido University, Faculty of Fisheries 1968). Research fishing north of 62°N has been conducted in late July and early August, probably after most sockeye salmon have entered their home river systems.

Mature sockeye salmon captured in offshore waters near Cape Navarin and the Gulf of Anadyr, both in the western Bering shelf area, are probably bound for the Tumana and Anadyr rivers.

Spawning migration, chum salmon

Chum salmon are more widely distributed throughout the shelf during the spawning migration than sockeye salmon, probably because chum salmon spawn in many more streams in Norton and Kotzebue Sounds (see Figs. 35-2 and 35-3).

In the southeastern shelf area along the approaches leading to Bristol Bay, chum salmon appear to be more abundant farther offshore than sockeye salmon—a phenomenon undoubtedly related to the size of the chum salmon populations returning to the river systems tributary to Bristol Bay. On the average, river systems that enter on the north side of upper Bristol Bay produce over 70 percent of the commercial chum salmon catch of the Bristol Bay area (ADF&G 1974). Chum salmon probably begin to separate according to the geographic location of the major chum salmon producing river systems on the north side of Bristol Bay when they are more than 200 km seaward of the river mouths.

Chum salmon apparently bound for the Yukon River, arctic coast, Gulf of Anadyr, and Norton and Kotzebue Sounds are more abundant in the western than in the eastern shelf areas. Some have been found to be moving eastward both north and south of St. Lawrence Island (Yonemori 1967); they were probably bound for the Yukon River, Norton Sound and Kotzebue Sound rivers, and rivers entering the Bering Sea along the eastern and western arctic coast. A few tagged very near the shelf edge in the southeastern Bering Sea (between 54-56°N and 170°W) were recovered on Hokkaido and Sakhalin Islands and in the eastern Kamchatka area, indicating westerly movement for these fish. Chum salmon captured in research gillnets in the western shelf area were moving northward in Anadyr Bay (Yonemori 1967) and were most likely bound for the rivers entering this bay.

Spawning migration, pink salmon

Pink salmon have been captured throughout the offshore areas of the shelf during spawning migration (Fig. 35-4). As with sockeye and chum salmon, the heaviest concentrations on the shelf appear to be related to the size of the population migrating to specific stream or river systems and the geographic location of these systems.

The Nushagak River system, the most important pink salmon producing river system in Bristol Bay,

drains into the north side of the bay. Between 80 and 90 percent of the total Bristol Bay catch of pink salmon is taken in the Nushagak fishing district (ADF&G 1974). The other major pink salmon producing river systems in Bristol Bay include the Togiak River system, which also enters the bay on the north side, and the Naknek and Kvichak river systems, which enter at the head of the bay. The commercial catch in the north side peninsula commercial fishing districts (see Table 35-1) indicates that relatively few pink salmon are produced in the streams along the north side of the Alaska Peninsula. Pink salmon, like chum salmon, are produced in greater abundance in river systems entering on the north side than on the south side of Bristol Bay. They are also more abundant farther offshore in the northward approaches to the bay than sockeye salmon (see Figs. 35-2 and 35-4).

Some pink salmon of Bristol Bay origin apparently migrate a considerable distance north in the central shelf area and then southeast to Bristol Bay. For example, a single pink salmon tagged at 62°N, 170°W in the summer of 1966 was recaptured in the Nushagak area of Bristol Bay (Yonemori 1967). This fish had actually by-passed the latitude of its home river system by more than 200 km and migrated more than 800 km in a southeast direction from point of tagging to point of recapture. What proportion of the pink salmon of Bristol Bay origin undertake this rather extensive migration across the shelf is unknown.

Many of the streams and river systems entering Norton Sound produce pink salmon. The distribution and abundance noted in the central Bering Sea shelf area suggest that they are probably bound for streams and rivers entering Norton Sound and Kotzebue Sound. Pink salmon captured in research gillnets of St. Lawrence Island were moving eastward (Yonemori 1967) and were probably destined for the streams of Norton and Kotzebue sounds. A few captured in research gillnets west and northwest of St. Lawrence Island (Hokkaido University, Faculty of Fisheries 1965 and 1968), probably moving north and northeastward, were also probably destined for Kotzebue and Norton sounds. Pink salmon captured in the Gulf of Anadyr were moving northward (Yonemori 1967) and most were probably destined for the Anadyr River.

Spawning migration, chinook salmon

Maturing chinook salmon have been captured throughout the Bering Sea shelf during spawning migration, but little information is available to verify the direction of migration at this time. Information on distribution and direction of migration for sock-

eye, chum, and pink salmon in the western, central, and southeastern shelf areas, respectively, suggests that they are primarily bound for rivers and streams of the Gulf of Anadyr, Norton and Kotzebue sounds, and the Alaska Peninsula and Bristol Bay (see Figs. 35-2, 35-3, and 35-4). If chinook salmon are responding similarly to the same environmental cues as sockeye, chum, and pink salmon to guide them during spawning migration, they can be expected to follow migration routes similar to those of these species. On this assumption, the probable directions of migration of chinook salmon captured in various areas of the Bering Sea shelf are shown in Fig. 35-5.

Chinook salmon were captured most frequently in Bristol Bay, where research fishing concentrated on the sockeye salmon runs to this area; chinook were generally captured in greater abundance, farther north, and farther offshore than sockeye (see Figs. 35-2 and 35-5). The distribution of chinook salmon is probably related to the size of the population migrating to specific river systems in Bristol Bay; they are produced in far greater abundance in streams entering the north side of Bristol Bay. Catches in fishing districts on the north side of Bristol Bay (i.e., Nushagak and Togiak bays) have amounted to more than 80 percent of the total Bristol Bay catch of this species. Chinook salmon are also abundant in the Kuskokwim River, north of Bristol Bay.

Spawning migration, coho salmon

Coho salmon have been captured at only a few locations on the Bering Sea shelf (see Fig. 35-6), primarily because there has been no exploratory fishing from late July to late August, when they would be most abundant on the shelf (see Table 35-3). They have been captured most frequently in the southeastern Bering Sea shelf area, where research fishing has concentrated on the sockeye salmon migration bound for the river and stream systems tributary to the Alaska Peninsula and Bristol Bay.

The same rationale used to conjecture the probable migration route of chinook salmon in the shelf area is used for that of coho salmon. There is little direct evidence, however, to verify these.

General seaward migration

The seaward migration of Pacific salmon begins in spring, when the juveniles enter the shelf area from their rivers or streams of origin. The seasonal timing of this migration is species and stock specific and variable according to annual differences in environmental conditions, such as time of ice breakup on lakes, streams, and rivers and the warming of these waters. Exploratory fishing for juvenile salmon (Na-

tional Marine Fisheries Service, Auke Bay Laboratory 1966 and 1967 data) and casual observations by biologists (Mike Nelson, Alaska Department of Fish and Game, personal communication) in Bristol Bay indicate that juvenile chinook salmon are earliest to enter the bay in mid- to late May, and sockeye, chum, pink, and coho salmon, in that order, enter from late May to late July.

After entry into the Bering Sea, juvenile salmon remain in the nearshore waters for varying lengths of time during the initial few months of seaward migration (Hartt et al. 1967b, Straty 1974, and Barton 1979). Observations in the Gulf of Alaska (Hartt et al. 1967a; Hartt and Dell 1976), along the south side of the Alaska Peninsula and eastern Aleutian Islands (Hartt and Dell 1976), and in the western North Pacific Ocean and Okhotsk Sea (French et al. 1976) indicate that coastal movement during the first few months of seaward migration is typical behavior for Pacific salmon throughout their range.

As the seaward migration of salmon progresses from early summer to late fall, the fish rapidly increase in size and ultimately leave the nearshore waters to move offshore to more pelagic regions (Straty 1974, Barton 1979). The seasonal timing of this offshore movement of juvenile salmon is species specific and variable according to annual differences in time of entry into the Bering Sea. For example, in years when ice breakup and warming of water occur early, offshore movement would be expected to occur earlier in the season than when those conditions occur later. The location and timing of offshore movement in a given year are probably a function of time of entry into the Bering Sea and growing conditions. The more rapidly juvenile salmon increase their size, the sooner they can be expected to move offshore. Information is only fragmentary on the shelf distribution of juvenile salmon after they leave the coastal waters.

Of the five species of juvenile Pacific salmon inhabiting the shelf, only the sockeye salmon of Bristol Bay have been studied sufficiently to describe in some detail their seaward migration through the southern shelf area. Information on the seaward migration of the other species of salmon across the shelf is fragmentary and largely the result of the sockeye salmon studies of Hartt et al. (1964, 1967a and b), Straty (1974), and Straty and Jaenicke (in press) and of finfish surveys conducted in Norton and Kotzebue Sounds (Barton 1979). Much of what was learned about the behavior of juvenile sockeye salmon during seaward migration through Bristol Bay probably applies, in varying degrees, to the other species of juvenile salmon.

Seaward migration, juvenile sockeye salmon

What is known of the distribution, migration routes, and ecology of juvenile sockeye salmon in the southeastern shelf area of Bristol Bay has been described in some detail by Straty (1974) and Straty and Jaenicke (in press). I summarize here what is known of the distribution and migration routes of juvenile sockeye salmon across the shelf area between late May and late September.

Juvenile sockeye salmon from all river systems entering Bristol Bay, including systems along the north side of the Alaska Peninsula, apparently follow the same seaward migration route regardless of the specific geographic location of their river systems of origin (i.e., whether the river enters on the north or south side of Bristol Bay: Straty 1974). This coastal route is along the north side of the Alaska Peninsula (Fig. 35-7). From late May to late September, juvenile sockeye salmon are most abundant between the coast and 48 km offshore. The fact that all stocks of juvenile sockeye salmon and apparently other species of juvenile salmon follow this seaward route suggests that they all respond to the same environmental conditions, which guide them seaward across the shelf and along the north side of Bristol Bay.

Virtually nothing is known of the distribution and seaward migration routes of juvenile sockeye salmon originating in rivers north and northwest of Bristol Bay (i.e., those entering Kuskokwim Bay, Norton and Kotzebue sounds, and the Gulf of Anadyr). The probable seaward migration routes of juvenile sockeye salmon from these river systems are illustrated in Fig. 35-7, based on the apparently typical movement of juvenile salmon along the coast throughout their range during the initial period of seaward migration. Similar to Bristol Bay sockeye salmon, these fish probably vacate the nearshore waters and move offshore after an undetermined time.

Juvenile sockeye salmon originating in the various rivers of Bristol Bay, and probably rivers in other areas as well, enter the sea at different times during late spring and early summer (Straty 1974, Rogers 1977). Sockeye salmon migrating from Bristol Bay lake systems with only one or two lakes enter Bristol Bay earlier and over a shorter period of time than those migrating from multi-lake systems. In addition, the distances juvenile sockeye salmon travel from lake outlet to Bristol Bay differ for each of the river systems. As a result, the individual major river stocks are somewhat separated during the initial few weeks of seaward migration. Differences in their age and size at the time of entry into Bristol Bay also contribute to this initial separation. The larger, age-2.0 sockeye salmon precede age-1.0 sockeye salmon

Fig. 35-7. Distribution of sockeye salmon during seaward migration, mid-May through September. (Adapted from Straty 1974, Straty and Haight 1979.)

in seaward migration. Because age-2.0 fish from a given river system are larger than age-1.0 fish, they probably migrate seaward faster. The result of the differences in fish size and time of entry into Bristol Bay is that juvenile sockeye salmon are distributed throughout most of the area encompassing the seaward migration route from late May through late July.

Although the major stocks of Bristol Bay sockeye salmon may be segregated during the initial period of seaward migration, they mix after the fish have been at sea a few months (Straty 1974). This most likely results from a decrease in the seaward migration rate of those fish that have entered the more seaward areas of Bristol Bay, where they apparently encounter more abundant food than in the inner Bay.

From late May and early June through early August in years when usual environmental conditions prevail (e.g., mainly sea temperature), the greatest biomass of juvenile sockeye salmon will be present along the coast of inner Bristol Bay northeast of 159°W (Straty 1974). Early seaward-migrating sockeye salmon will be more abundant farther seaward in this area than later entrants into the bay. After early August, the greatest biomass of juvenile sockeye salmon apparently is in the outer bay or seaward of 159°W. Little is known about the distribution and migration route of juvenile sockeye salmon in Bristol Bay after late September, when most studies ended.

The seasonal distribution and abundance of juvenile sockeye salmon in the shelf area of Bristol Bay may show considerable annual variation. For example, in 1971, a year characterized by anomalously cold sea temperatures from spring through fall (Straty and Jaenicke, in press), juvenile sockeye salmon were virtually absent in outer Bristol Bay in early July, whereas they were abundant in this area in 1967, a year with warm sea temperatures, in early to mid-June. Such annual variations in the seasonal distribution of juvenile salmon can probably be expected in other areas of the shelf as well, and probably result in annual variations in the length of time juvenile salmon are residents of the shelf and the time of entry into the North Pacific Ocean.

Finally, juvenile sockeye salmon were found to be schooled and most abundant in the upper 5 m of water (Straty 1974) during seaward migration. They were most abundant at night in the upper 1 m of water and during daylight in a depth of 2 m.

Seaward migration, juvenile chum salmon and pink salmon

Except in Bristol Bay and Norton Sound, little is known about the shelf distribution of juvenile chum and pink salmon. In Bristol Bay, both species have been captured during research fishing at various times between early summer and early fall in the area encompassing the seaward migration route of the juvenile sockeye salmon (see Fig. 35-7) (National Marine Fisheries Service, Auke Bay Laboratory 1966 and 1969-72 data). Few juvenile chum and pink salmon have been captured outside this area. In Norton Sound (primarily Golovnin Bay), juvenile chum and pink salmon were captured in beach seines in the nearshore littoral areas between the onset of ice breakup 9 June, and 9 July 1977 (Barton 1979).

In Bristol Bay, small numbers of chum salmon were captured by purse seine in the coastal waters as early as mid-June 1967. However, chum salmon did not become abundant in these waters until after mid-July in 1967 and 1969. Juvenile chum salmon remained abundant along the southeastern coast of Bristol Bay seaward of 159°W through August and until at least mid-September in 1969 and 1970 (National Marine Fisheries Service, Auke Bay Laboratory 1969 and 1970 data). Juvenile pink salmon have not been captured in Bristol Bay in the area encompassing the seaward migration route of the juvenile sockeye salmon until after late June (National Marine Fisheries Service, Auke Bay Laboratory 1966 data). They have been captured primarily in the coastal areas of inner Bristol Bay northeast of 159°W. Their abundance in this area increased from

late June through mid-August of 1969. Presumably, juvenile pink salmon did not reach the coastal areas of outer Bristol Bay seaward of 159°W until late August and September. A few were captured near Port Moller in outer Bristol Bay (at $56^{\circ}12'\text{N}$, $161^{\circ}00'\text{W}$) on 25 August 1969. These were the only juvenile pink salmon captured seaward of 159°W during the various years of intensive research fishing in the area by Auke Bay Laboratory.

In Norton Sound, juvenile chum and pink salmon captured in the nearshore littoral areas during June and early July were less than half the size (32-59 mm mean fork length) of those captured between late June and mid-September in Bristol Bay. No juvenile chum or pink salmon of the larger size captured in Bristol Bay were taken in the littoral or offshore areas of Norton Sound after early July 1977. Barton (1979) concluded that juvenile chum and pink salmon vacated the nearshore littoral areas of Norton Sound and moved offshore to more pelagic areas after early July. But I have found gillnets and tow nets, used to sample juvenile salmon in these areas, to be much less efficient than purse seines or round-haul seines for sampling juvenile salmonids in Bristol Bay. Small juvenile chum and pink salmon present in the offshore areas of Norton Sound after early July probably missed being caught in the gill nets and tow nets, or else avoided them. After the juveniles moved offshore, they would have been larger in size, more dispersed, and more difficult to sample. It seems unlikely that small juvenile salmon of the size reported by Barton (1979) would have migrated into distant, deep offshore waters. This behavior is inconsistent with behavior observed for juvenile Pacific salmon throughout their range in North America.

In Bristol Bay, no juvenile chum or pink salmon of the small size sampled in Norton Sound were captured during research fishing by Auke Bay Laboratory. However, in Bristol Bay, littoral areas less than 16-18 m deep were not sampled for juvenile salmon. Since the mesh size of the purse seine used to sample juvenile salmon was not capable of retaining fish smaller than 60-70 mm, the smaller-sized juvenile chum and pink salmon may have been present in the littoral water of Bristol Bay during early June but not sampled. After increasing in size and moving progressively offshore, they would have become available to the fishing gear used.

Both chum and pink salmon enter the estuaries shortly after emerging from the gravel (see Table 35-2). They are considerably smaller (25-30 mm) than juvenile sockeye, chinook, and coho salmon; however, few juveniles less than 25-30 mm long have

been captured in Bristol Bay, probably because they were not sampled in the nearshore littoral areas where they are likely to occur. Juvenile pink salmon in this size range have been seen milling around cannery docks in Bristol Bay in early June. Juvenile chum and pink salmon during seaward migration in the shelf area probably move offshore progressively as their size increases.

Seaward migration, juvenile chinook salmon and coho salmon

Juvenile chinook and coho salmon have also been captured with juvenile sockeye salmon during research fishing in the shelf area of Bristol Bay. The distribution of juvenile chinook and coho salmon is similar to that of other salmonids in Bristol Bay: they are most abundant along the southeastern coast (see Fig. 35-7). A few juvenile coho salmon were captured during finfish surveys of Golovnin Bay in Norton Sound (Barton 1979), but neither their distribution nor movements were noted.

In Bristol Bay, only a few juvenile chinook salmon were captured during research fishing by the Auke Bay Laboratory in 1966, 1967, 1969, and 1970. These fish were caught in early June 1966 in the inner bay northeast of 159°W . None have been captured in the shelf area of the Bay beyond June. Between June and August, a few were taken during research fishing farther seaward in the eastern Bering Sea (Hartt and Dell 1976).

The absence of juvenile chinook salmon in the shelf areas of Bristol Bay after early June indicates either that the majority of the chinook salmon had entered the bay before intensive fishing began or that for some unexplained reason they were missed by the fishing gear. In the Salcha River drainage of the Yukon River, most chinook salmon migrate downstream between the second half of May and early June (Trasky 1974). Chinook salmon from Bristol Bay river systems probably also migrate seaward during May and early June. The absence of juvenile chinook salmon in the shelf area of Bristol Bay beyond June also implies that these fish migrate seaward much faster than the other salmon species, which enter the bay later in the spring and summer than the juvenile chinook salmon and remain in the shelf area at least through mid-September. Juvenile pink salmon probably remain in the shelf area into October.

According to the time of their appearance and the magnitude of their numbers as indicated by fish catches made in the coastal waters of Bristol Bay, juvenile coho salmon are the latest species of Pacific salmon to enter the shelf. Although they have been

captured along the southeastern coast of Bristol Bay as early as mid-June, research fishing indicates that they are not abundant until late June or early July (National Marine Fisheries Service, Auke Bay Laboratory 1966, 1967, 1969, and 1970 data); they remain abundant throughout July and August. Some juvenile coho salmon were still present on the shelf in mid-September of 1970.

Synopsis: Timing of migration and distribution

Exploratory and research fishing and tagging experiments have shown that sockeye, chum, pink, chinook, and coho salmon are widely distributed throughout the Bering Sea shelf during their spawning migrations. They are more abundant in offshore waters than in coastal waters until they reach the vicinity of their home river or stream systems. Adult salmon are most abundant in the shelf area between mid-May and early September; the peak of abundance differs somewhat for each species (see Table 35-3).

Adult sockeye, chum, and pink salmon, and probably also chinook and coho, appear segregated to varying degrees according to the general geographic location of their home river or stream systems at the time they enter the Bering Sea shelf. Salmon bound for the Yukon River, Norton and Kotzebue sounds, and the arctic coast appear to be more prevalent in the central shelf area. Those bound for the rivers of the Gulf of Anadyr appear more prevalent in the western shelf area, chum salmon in the western and central shelf areas. Salmon bound for rivers and streams in Bristol and Kuskokwim Bays and along the Alaska Peninsula appear more abundant in the eastern shelf area. The general similarities in the shelf distributions of sockeye, chum, and pink salmon bound for streams or rivers in such specific geographic areas as Bristol Bay suggest that these species, and probably chinook and coho as well, are responding inherently to the same environmental cues to guide them across the shelf to their rivers or streams of origin. Acceptance of this hypothesis implies that the spawning migration of salmon in the shelf area is directed along broad but specific routes that must be similar for all species.

During the spawning migration when salmon proceed across the shelf in the general direction of their rivers of origin, the different stocks progressively segregate according to the specific location of their home river or stream systems, such as the Nushagak River in Bristol Bay. Sockeye salmon stocks of Bristol Bay origin begin to segregate when the fish are still in the offshore waters, as much as 200 km from the mouths of home river systems (Straty 1975).

The timing of the seaward migration of juvenile Pacific salmon across the shelf is similar to that of the adult spawning migration (i.e., juvenile salmon are most abundant on the shelf between early to mid-May and late September). Although both juvenile and adult salmon are present on the shelf at the same time, their distributions and migration routes are quite different. Adult salmon appear to remain in the offshore waters of the shelf until they are near the coastal area or the estuary containing their home river or stream. Juvenile sockeye, chum, pink, and coho salmon, however, appear to move along the coast during the initial two months or so of seaward migration. Therefore, although the shelf is occupied by both juvenile and adult salmon at the same time of the year, they remain separated except, perhaps, near the mouths of their home rivers or streams. Little information is available on the distribution of juvenile chinook salmon in the shelf area. The small number caught in inner Bristol Bay suggests that the seaward migration route of this species is also along the coast.

The difference in their migration rates keeps juvenile salmon largely separate from the adults during seaward migration. For example, the majority of the adult sockeye salmon bound for Bristol Bay river systems cross the shelf in about 15 days (see Table 35-3); juvenile sockeye salmon, and probably juvenile chum, pink, and coho, appear to be confined to the coastal waters for at least 60 days during seaward migration. As a result, adult salmon have crossed the shelf offshore well before the juvenile salmon begin to move into that area.

Observations in Bristol Bay, Norton Sound, and the Gulf of Alaska indicate that juvenile salmon eventually move from the coastal waters farther offshore to more pelagic waters as their size increases. In Bristol Bay, the larger-sized juvenile sockeye salmon seem to be the first to move offshore (Ogi 1973; National Marine Fisheries Service, Auke Bay Laboratory 1967, 1969, and 1970 data). If the time of offshore movement of juvenile salmon depends on their size, it can be expected to vary annually and seasonally, as a result of variations in environmental conditions such as sea temperature and food abundance, which influence growth.

ENVIRONMENTAL INFLUENCES ON THE MOVEMENTS AND DYNAMICS OF PACIFIC SALMON ON THE BERING SEA SHELF

Fishery scientists generally agree that environmental conditions such as salinity, currents, food abundance, and water temperature influence the growth, movement, distribution, and survival of

salmon. To make a comprehensive synopsis of the movements and dynamics of Pacific salmon, one must understand how environmental conditions and their variability affect the behavior of salmon residing on the shelf.

Considerable research has been conducted in the laboratory on the influence of the environment on the physiology and behavior of salmon, but application of results to explain the behavior of salmon at sea has been limited. Numerous studies conducted in Bristol Bay have focused on the apparent responses of sockeye salmon to certain environmental conditions and their variability (Ogi 1973, Straty 1974, Nishiyama 1974, Fujii et al. 1974, Fujii 1975, Straty and Haight 1979, and Straty and Jaenicke, in press). The results of these studies and the inferences made from them concerning probable movements and dynamics of Pacific salmon of the shelf are summarized below.

Influence of temperature on the distribution, growth, and survival of Pacific salmon

Environmental temperature has profound effects on the metabolism and, consequently, the life processes of fish. Temperature limits the metabolic rate within which the fish are free to perform, regulates their rate of growth and development, and acts as a directive element resulting in aggregation of fish within thermal ranges or in movement of fish to new environmental conditions (Brett 1956). The range of tolerable and preferred sea-surface temperatures is different for each species of salmon in the Bering Sea and North Pacific Ocean, and these ranges change from spring through fall (Table 35-4, Manzer et al. 1965). Sockeye and chum salmon preferred the lowest sea-surface temperature range; pink salmon, the intermediate range; and chinook and coho salmon, the highest range. From May through July and August, all salmon species preferred increasingly

higher temperatures; in September, they again preferred colder waters. In spring and early summer, maturing salmon were associated with slightly colder sea-surface temperatures than immature salmon. In late summer and fall, maturing salmon tended to be associated with warmer sea-surface temperatures.

The studies of Manzer et al. (1965) lend credence to the hypothesis, first advanced by Kaganovsky (1949) with respect to pink salmon, that in general, salmon avoid waters of near-freezing surface temperatures and vacate the Bering Sea and northwestern Pacific Ocean in winter to move southward and eastward where water temperatures are more favorable. Winter sea-surface temperatures over much of the Bering Sea and northwestern Pacific Ocean are less than 3 C (Laviolette and Seim 1969). Fujii (1975) concluded that sockeye salmon of western Alaska origin are prevented from entering the Bering Sea through passes in the Aleutian Islands until low temperatures and high salinities of surface water in the passes disappear. Manzer et al. (1965) speculated that in years when water temperatures are relatively high, sockeye and chum salmon may be present in the southern Bering Sea throughout the winter months because of their lower preferred temperature range. Sockeye salmon were, in fact, captured by gillnetting in January in the central Bering Sea between 180° and 175°E (Hunter and Larkins 1963).

If sea temperature is important in regulating the seasonal distribution of Pacific salmon, the proportion of juvenile and immature salmon overwintering in the Bering Sea should be correlated to the amount winter sea temperatures exceed the species' minimum preferred temperatures.

Little information is available on the winter distribution of juvenile and maturing salmon on the shelf. Since salmon are present in the Bering Sea in winter, it is possible that they may enter the shelf, particularly in years characterized by anomalously warm winter sea temperatures.

The depth distribution of some species of Pacific salmon may be restricted by sea temperature (INPFC 1963, Manzer 1964). In the early part of the summer season when a thermocline had not developed, salmon were scattered vertically; after the thermocline had developed, salmon tended to concentrate in the surface layer (INPFC 1963). A well-produced thermocline appears to be a physical barrier through which sockeye salmon do not migrate. Chum salmon, however, reacted this way to the thermocline only during hours of darkness. It is not known to what extent vertical movements of salmon are restricted by development of a thermocline on the shelf during summer.

TABLE 35-4

Probable tolerable and preferred sea-surface temperatures for Pacific salmon (from Manzer et al. 1965)

Species of salmon	Tolerable range (C)	Preferred range (C)	Reference months for preferred range
Sockeye	1-15	2, 3-9	May, September
Chum	1-15	2, 3-11	May, September
Pink	3-15	4-11	May, June
Coho	5-15	7-12	May, June, July
Chinook	2-13	7-10	July, August, September

In addition to restricting the seasonal and vertical movements of salmon, sea temperature may also influence the width of the seaward migration route across the shelf (Straty 1974). Coastal waters on the southeastern side of Bristol Bay are generally a few degrees warmer than adjacent waters offshore (i.e., isotherms parallel the coast). During seaward migration through this area, juvenile salmon are afforded a range of temperatures that may include those best suited to their thermal requirements at the time. Juvenile sockeye salmon appear to avoid the colder offshore waters during seaward migration, particularly in years (like 1971) when anomalously cold sea temperatures prevail (Straty 1974). In 1971, juvenile sockeye salmon were captured during research fishing by National Marine Fisheries Service only at locations nearest the coast and in the shallowest water that could be fished (National Marine Fisheries Service, Auke Bay Laboratory, 1971, unpublished data).

Sea temperature not only restricts the seasonal movements of salmon on the shelf but also modifies their movements through its influence on growth and swimming speed (Straty 1974, Straty and Haight 1979, and Straty and Jaenicke, in press). Year-to-year sea temperature differences of 6 C between May and December have occurred in Bristol Bay (Fig. 35-8).

Fig. 35-8. Average monthly sea-surface temperatures at 57°N 163°W in Bristol Bay during May-December 1967 and 1971. Vertical line represents time of peak migration of sockeye salmon from Bristol Bay rivers. (From Straty and Jaenicke, in press.)

Laboratory experiments have shown that changes in sea temperature of this magnitude profoundly affect the swimming speed of young sockeye and coho salmon (Brett et al. 1958). As temperature increased from 2 C to 10 C, the swimming speed of juvenile sockeye salmon increased over 80 percent. Juvenile

coho salmon showed a similar increase in swimming speed between 4 C and about 12 C. Annual variations in sea temperature of the magnitude shown in Fig. 35-8 should cause significant annual differences in swimming speed and hence in the seaward migration rate of juvenile salmon across the shelf. As a result, annual differences can be expected in the distribution of juvenile salmon on the shelf and in their time of entry into the offshelf regions of the Bering Sea and the North Pacific Ocean. Annual differences in the seasonal distribution of juvenile sockeye salmon have been observed in Bristol Bay and have been attributed to these extreme variations in sea temperature (Straty 1974, Straty and Haight 1979).

Sea temperature may also influence the rate of seaward migration³ and possibly survival, through its influence on growth (Straty 1974, Straty and Haight 1979, and Straty and Jaenicke, in press). Young sockeye salmon do not grow at temperatures below 4 C (Donaldson and Foster 1940). Low temperatures apparently reduce the rate of digestion which in turn reduces consumption and growth (Brett and Higgs 1970). Growth and food utilization were highest at approximately 10 C. Juvenile salmon migrating seaward through Bristol Bay experienced significantly different growing seasons⁴ in 1967 and 1971 (see Fig. 35-8). The growth pattern of scales of sockeye salmon from several river stocks of Bristol Bay that migrated seaward in 1967 were significantly different from those that migrated in 1971 (Straty and Jaenicke, in press).

Annual variations in the size of seaward-migrating salmon may be important for their survival. The survival rate of juvenile salmon apparently increases with increased size (Foerster 1954, Ricker 1962, Manzer 1972, Pella and Jaenicke 1978); therefore, environmental conditions such as warm sea temperature and abundant food, which favor rapid growth, are conducive to high early marine survival of juvenile salmon. A rapid increase in size (1) reduces the time juvenile salmon are subjected to size-dependent predation by fish and marine birds and mammals; (2) increases the rate of seaward migration, resulting in earlier entrance of juvenile salmon into the North Pacific Ocean; and (3) changes the food habits of juvenile salmon so that they are not competing for food items exploited by other fish and marine birds.

³ Swimming speed is a function of water temperature and fish length.

⁴ Growing season, which has been defined as the degree and duration of sea temperatures exceeding 4 C, influences the amount of growth a juvenile salmon attains during the first year of ocean life (Straty and Jaenicke, in press).

TABLE 35-5

Rate of growth per day and increment of growth of age-2.2 Bristol Bay sockeye salmon
in 1967 and 1971
(data from Nishiyama 1974.)

Year	Sea surface temperature (°C)	Sex	Body weight (g)			Rate of growth per day (%)	Increment of growth (g)
			Initial	End	Mean		
1967	8.5±1.02	Female	2106	2412	2259	0.753	17.00
		Male	2426	2750	2588	0.696	18.00
1971	4.6±0.74	Female	2086	2156	2121	0.471	9.98
		Male	2343	2413	2378	0.438	10.41

Since juvenile salmon reside longer on the shelf than maturing salmon, annual variations in sea temperature may be expected to have a more pronounced and measurable effect on the distribution, growth, and abundance of juvenile salmon than on maturing salmon. However, sea temperature apparently also influences the growth rate of maturing Bristol Bay sockeye salmon during their migration across the shelf (Table 35-5, Nishiyama 1974). The rate of growth per day and increment of growth in 1967, a year with warm sea temperatures (see Fig. 35-8), was 60 percent greater than in 1971, a year with cold sea temperatures.

In conclusion, annual variations in sea temperature of the magnitude occurring in Bristol Bay significantly affect the distribution and growth of maturing and juvenile Pacific salmon while they are residents of the shelf. Annual variations in sea temperature may also indirectly influence the survival rate of juvenile salmon by influencing growth rate.

Influence of abundance and type of food on growth and distribution

In addition to being a function of environmental temperature, growth in salmon is a function of the amount and type of food consumed. The potential energy available to salmon for maintenance and growth is different for each type of food organism eaten (Nishiyama 1974). Annual, seasonal, and areal variations in abundance and type of food eaten by salmon residing on the shelf have been reported by Dell (1963), Nishiyama (1974), Straty (1974), Carlson (1976), and Straty and Jaenicke (in press). In Bristol Bay, variations in food type and abundance may result in corresponding variations in the distribution and growth and, therefore, survival of juvenile sockeye salmon.

In the shelf area of Bristol Bay, zooplankton (the food of juvenile sockeye salmon) were considerably more abundant along the migration route of juvenile sockeye salmon (see Fig. 35-7) seaward of 159°W (Straty and Jaenicke, in press) than toward the inner bay. This difference was apparent from June through August. In addition, the coastal water of the bay (less than 37 m deep) contained significantly fewer zooplankters than the offshore waters (deeper than 37 m). The fullness of the stomachs of juvenile sockeye salmon sampled in these areas reflected these differences in zooplankton abundance. Juvenile sockeye salmon generally consumed more food farther seaward and farther offshore.

The response of juvenile sockeye salmon to greater food abundance in the more seaward portion of the Bristol Bay shelf area was clearly evident in the size of the fish and growth patterns of their scales. Juvenile sockeye salmon seaward of 159°W were larger than those in the inner bay northeast of 159°W. The appearance of widely-spaced marine-type circuli on the scales of these fish also showed that initial marine growth did not occur or accelerate until the juvenile sockeye salmon had entered outer Bristol Bay seaward of 159°W (Straty 1974). Juvenile sockeye salmon usually reach the outer bay after 4-6 weeks, although this may vary annually. Juveniles appear to move more slowly seaward and spend more time feeding in the outer bay where food is abundant than in the inner bay.

Food of juvenile sockeye salmon in the outer bay was not only more abundant but also larger, a fact which may contribute to the increased growth rate of juvenile sockeye salmon observed in outer Bristol Bay (Straty and Jaenicke, in press).

Variations in food type and abundance must also influence the distribution and growth of maturing

salmon crossing the shelf, but such relationships have not been described. The annual differences in the rate of growth per day shown for maturing Bristol Bay sockeye salmon during late June and early July of 1967 and 1971 (see Table 35-5) (Nishiyama 1974) are probably due to the combined effects of annual variations in sea temperature and the type and abundance of food consumed.

Influence of salinity

As shown in Fig. 35-7, the seaward migration route of juvenile sockeye salmon from major Bristol Bay river systems is from the river mouths to the southeastern side of inner Bristol Bay. This movement passes through the area of the most pronounced salinity gradients (Straty 1974, Straty and Jaenicke, in press) and supports the experimental conclusion of McInerney (1964) that the juveniles of the five species of Pacific salmon occurring in the shelf area were able to use estuarial salinity gradients as one of the directive cues in their seaward migration. Juvenile salmon displayed a temporal progression of salinity preference changes which parallels the salinity gradients typical of river outflows experienced by young salmon on their way to the ocean. The terminal preference is the salinity of the open ocean.

The apparent response of juvenile sockeye salmon to salinity gradients in Bristol Bay is best exemplified by the movement of fish migrating seaward from the Wood River system. These fish left the Wood River on the northern side of inner Bristol Bay and migrated across the bay to the southeastern side (Straty and Jaenicke, in press) experiencing a horizontal change in salinity of 30‰ in a distance of 120 km. This is their most direct route through the turbid water of the inner bay to the less turbid waters of higher salinity in the outer bay. If salinity gradients are the principal environmental condition directing the initial movement of juvenile salmon after they enter the shelf, environmental events, such as prolonged periods of high wind, which cause shifts in the position of river outflows may produce short-term changes in distributions.

Salinity may also influence the distribution of maturing Bristol Bay sockeye salmon during migration across the shelf (see Fig. 35-2). Fujii (1975) has implied that the spawning migration route of sockeye salmon in this area is related to the position of water of high salinity flowing eastward into Bristol Bay. This wedge of high-salinity water extends into inner Bristol Bay, northeast of 58°N, 158°W (Straty 1974) and is distinguished from coastal water in the outer bay by the 32.0‰ isohaline (Fujii, 1975). This water mass is found between 56-57°N latitude in

outer Bristol Bay but varies in position from year to year. If maturing sockeye salmon prefer or are confined to this water mass, the position of their main migration route may also vary annually.

SYNOPSIS: INADEQUACY OF EXISTING KNOWLEDGE OF PACIFIC SALMON OF THE BERING SEA SHELF

Deficiencies in our knowledge of the distribution, movements, and abundance of Pacific salmon while residents of the shelf are apparent. Intensive investigations of sockeye salmon in the southeastern shelf and adjacent river system have provided a more complete description of salmon distribution, movements, and abundance in this area than in other areas of the Bering Sea shelf. As a by-product of these investigations, some information has also been obtained on the timing of migrations and distribution of maturing and juvenile chum, pink, chinook, and coho salmon. Information about the distribution and movement of salmon in the central, western, and northern shelf areas is based on few studies or has been inferred from the results of the salmon investigations in the southeastern shelf area.

The species composition, distribution, abundance, and movements of salmon on the shelf between November and April are largely unknown. Juvenile chinook salmon have been caught in bottom trawls near the shelf edge in mid-winter, suggesting that other salmon species may also occur here during the winter, perhaps in years when winter sea temperature exceeds the minimum preferred temperatures of the species.

In Bristol Bay, where the only intensive studies of juvenile salmon have been conducted, virtually nothing is known of their distribution and movements after late September. Juvenile salmon are probably present on the shelf at this time, particularly the smaller pink salmon and the seaward-migrating coho salmon. Juvenile salmon migrating from rivers and streams in the northern shelf area, e.g., Kotzebue and Norton sounds, may be present in the coastal and offshore areas of the shelf even later than those migrating seaward from Bristol Bay rivers. Because this is largely speculation, additional study is required to verify such occurrences.

Since most investigations of salmon have been conducted on the shelf between May and September, our knowledge of the occurrence and movements of salmon on the shelf before and after this period is extremely deficient.

Distribution of maturing salmon

Because research fishing and tagging studies have taken place in mid-June and July when maturing sockeye salmon and, to a lesser extent, chum salmon of Bristol Bay origin were abundant in the southeastern shelf, their general distribution, movements, and the timing of their migration have been well defined. However, significantly less is known of the distribution and movements of maturing pink, chinook, and coho salmon in the southeastern shelf area.

Even less is known about the distribution and movements of maturing salmon of all species in the central, western, and northern shelf areas. Few tagging experiments have been conducted on the shelf west of 168°W and north of 59°N. As a result, knowledge of the direction of movement of a salmon species in a given area may be based on the recapture of a single tagged fish. Additional systematic research fishing and salmon tagging from early May to early October on the shelf west of 168°W and between the shelf edge north of 56°N and 66°N are required to define more precisely the distribution, movements, and origin of maturing Pacific salmon in these areas.

Distribution of juvenile salmon

The only salmon species of Bristol Bay origin whose juveniles have been investigated to any extent is the sockeye salmon. Studies of the timing of migration, distribution, and abundance of juvenile sockeye salmon have resulted in some information on the distribution of juvenile chum, pink, and coho salmon in the southeastern shelf area. Because these investigations were conducted between early June and late September, little has been learned of the distribution and movements of juvenile chinook salmon in this area. Chinook salmon apparently enter the shelf earlier than early June, and few have been captured in Bristol Bay. Little is known of the distribution and movements of juvenile pink salmon seaward of 159°W in Bristol Bay and after early August. Juvenile pink salmon seem to migrate seaward at a slower rate than the other salmon species because of their smaller size during the initial period of seaward migration; few pink salmon juveniles were captured before investigations ended in late September. The distribution and movements of all species of juvenile salmon in the southeastern shelf area beyond late September are unknown.

Virtually nothing is known about the shelf distribution and time of migration of juvenile salmon originating in rivers and streams north and west of Bristol Bay. On the basis of the observed distribution of juvenile salmon in Bristol Bay and throughout their

range in the North Pacific Ocean, it seems safe to assume that juvenile salmon from streams north and west of Bristol Bay will initially migrate seaward along the coast. These fish would also be expected to move offshore as their size increases, probably well before they cross the shelf. A significant population of age-0 salmon should be in the offshore waters of the shelf in late summer and fall. Verification of such an occurrence awaits the results of additional research.

One final point concerning the seaward movements of juvenile salmon is the direction of movement of those fish originating in rivers north of Bering Strait (e.g., in Kotzebue Sound and along the Siberian and North American arctic coasts). Do they migrate seaward initially along the coast on the east or west side of the shelf or through the offshore waters? The answer to this question also awaits the results of additional research.

Estimates of the abundance

Estimates of the annual abundance of maturing sockeye salmon (catch plus the spawning population) returning to Bristol Bay rivers and some rivers along the north side of the Alaska Peninsula represent the most accurate and complete population statistics for any salmon-producing area of the shelf. Similar estimates for other species of maturing salmon and rivers and streams in most other areas of the shelf are unavailable. For many of these other areas only annual estimates of the commercial and/or subsistence catch of each species of salmon are available. Estimates of the annual abundance of each species of spawning salmon in these areas, which may at times exceed the commercial and subsistence catches, are limited to a few key streams or are largely unavailable. As a result, the reliability of estimates of the annual abundance and biomass of maturing salmon crossing the shelf can be seriously questioned.

The reliability of similar estimates for juvenile salmon is also open to question, because annual estimates of the abundance of juvenile salmon have only been made for Bristol Bay and the northern side of the Alaska Peninsula shelf areas (Stern et al. 1976, Rogers 1977).

Variability in environmental conditions

Additional research to better define the distribution and movements of salmon on the shelf must provide a better understanding of salmon distribution related to variations in the physical and biological features of this area. Subjects that should receive special attention include: (1) the relationship between salmon distribution and growth and annual

fluctuations in the abundance and distribution of preferred food types, e.g., euphausiids, fish larvae; (2) identification of the sources and magnitude of food competition between juvenile salmon and other pelagic forms, including other salmonids, and possible effects of food competition on their growth; (3) the effect of the passage of a large population of maturing salmon on zooplankton abundance in the offshore waters and possible influence on juvenile salmon growth; (4) the effect on the available food supply and growth of late-migrating juvenile salmon of the passage of an abundant stock of early-migrating juvenile salmon; and (5) the effect of annual variations in sea temperature on the distribution and growth of salmon.

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Finfish and the Environment

36

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ABSTRACT

Characteristic features of the eastern Bering Sea shelf ecosystem are summarized and an annual chronology of monthly mean environmental conditions and events as well as distributions and spawning areas of halibut, pollock, herring, and yellowfin sole are presented for the purpose of providing a synthesis of the foregoing chapters of this section and indicating some aspects of the ecosystem dynamics of the area. The inadequacies of existing information on species behavior, multispecies interactions, and environmental relations are discussed and a number of specific fisheries oceanography studies are proposed. It is suggested that a more complete listing of sequences of resource-environment events be attempted and more multidisciplinary studies be conducted. When conditions, processes, and interactions are understood, the area should be considered as a potential site for shelf-ranching, i.e., selective protection, development, and harvesting of specific resources.

INTRODUCTION

The general intention of the chapters on halibut, herring, pollock, yellowfin sole, and salmon is to present a brief review of the biology of the different species of fish, discuss real or apparent environmental relations, and list future research requirements. As indicated in the first chapter of this section, this information was largely derived from studies designed to assess specific commercial stocks for management purposes. The original purpose of this chapter was merely to summarize the fish distributions in order to place them in clearer focus and, by inspecting temporal and spatial separations and overlaps, to show the potential value of multispecies and ecosystem

studies over present approaches to fisheries management and oil pollution studies. However, it became apparent that although annual NWAFC and other research cruises, usually in late spring and early summer, may provide adequate data for resource management and even inferences pertaining to warm and cold years, they afforded few specific environmental indices, and information on fish distributions during other parts of the year was fragmentary. Because in order to assess potential dangers from oil spills we need to know year-round distributions and activities of living marine resources, we have constructed an annual chronology of monthly conditions and events which, while reflecting the paucity of knowledge of the effects of the environment on fish, at the same time gives some insight toward potential understanding of relationships. And, in this vein, we have also posed sixteen questions that could provide the basis for various fisheries oceanography studies. Unlike some other sections of this volume in which investigators are reporting results of specific studies planned, approved, and funded by OCSEAP (i.e., physical oceanography and fisheries biology, which have been conducted independently), here we have attempted to synthesize fragmentary pieces into a dynamic system. If we are to advance our understanding of the area, the eastern Bering Sea shelf should be viewed as a system and all studies should be considered as merely components of the whole.

MONTHLY MEAN CONDITIONS AND EVENTS

There are a number of characteristic features of the eastern Bering Sea ecosystem that make it different from most other important fisheries areas (Table 36-1), but discussions will be limited to monthly mean environmental conditions and movements of halibut, herring, pollock, yellowfin sole, and salmon with the intent of supplementing information already presented.

Ice information presented stems largely from Potocky (1975) and NESS satellite imagery. Oceanographic conditions are derived from monthly summaries compiled at NWAFC from acquired and extant NODC geofile data (see Ingraham and Section I of this volume), in which data from October to March are extremely limited. Fisheries information in the form of schematic diagrams has been extracted from the foregoing chapters and discussions with authors.

January

Air temperatures range from near 0 C near the Alaska Peninsula to lower than -15 C near Bering Strait. The leading edge of drift ice extends over 300 km seaward trending northwest-southeast from the mid-Alaska Peninsula to Cape Navarin, passing between St. Matthew Island and the Pribilofs; driven by mean northerly winds, it advances 3-5 km/d. Water temperatures near -1.8 C (the minimum temperature possible at existing salinities) extend from inside the ice edge shoreward and isothermal and isohaline conditions occur under the ice; within 100 km seaward of the ice edge bottom temperatures may be several degrees higher because complete overturn usually occurs only under the ice. Bottom temperatures at the shelf edge and upper slope, and particularly southward of the ice edge in the southern part of the shelf, are 3-4 C, thus 5-6 C higher than in the vast area under the ice.

At this time most fish have retreated seaward and southward, aggregating over the outer shelf and upper slope between the Pribilof Islands and Alaska Peninsula: herring largely northward of the islands (but also to the northwest and southeast) over the outer shelf at 100-200 m; pollock and yellowfin sole at similar depths south of the islands but pollock also extending out over the slope and basin; and halibut seaward of the shelf edge and down the slope (Fig. 36-1). Halibut spawn, and eggs released at depths of 300-500 m rise slowly in the water column, larval stages eventually seeking the shallow shelf areas required for survival. The conditions, behavior, and migrations of herring, yellowfin sole, and other fish that may winter in coastal areas are unknown.

February

Environmental conditions are largely similar to those of January except for slightly lower air temperatures. The seaward advance of the ice edge continues, except in the area north of the Alaska Peninsula where the northeastward onshelf movement of warm water retards ice formation; however, northwest of this area an advance of 1-3 km/d occurs and ice extends seaward of the Pribilof Islands. Under appropriate air, sea, and wind conditions the ice edge may advance rapidly—thin surface ice may spread 50-100 or more km in a day and then either thicken or break up and melt, depending on subsequent weather conditions.

Halibut spawning is largely completed and eggs are dispersed by existing currents which are poorly defined and seemingly variable and complex (Hansen 1978). The existence of large and small eddies in the southwestern part of the shelf may serve to keep eggs and larvae in this area, but may also make them vulnerable longer to predation by the massive fish biomass. As a result of a special study aboard the *Miller Freeman* in 1978, Waldron (1978) indicated that pollock spawning may begin in this area in late February; unlike halibut eggs, pollock eggs are concentrated in the surface layer. Apparently there is little change in the distribution of herring, pollock, and yellowfin sole except perhaps a slight southward movement in the northern slope areas as water temperatures progressively decrease below 3 C.

Figure 36-1. Schematic diagram of January-February distribution of halibut, herring, pollock, and yellowfin sole. (Solid squares indicate spawning areas; long dashes indicate probable extension of shelf range; and broad line indicates ice edge.)

TABLE 36-1

Characteristic features of the eastern Bering Sea shelf ecosystem

Characteristic features	Consequences
<i>Physical features</i>	
Large continental shelf	High standing stocks of biota High fish production Large food resources for mammals
High latitude area	Nutrient replenishment with seasonal turnover Environmental distribution limits for many species Large seasonal changes Seasonal presence of ice Accumulation of generations
Large occasional changes	Seasonally changing growth Seasonal migrations Possibility of large anomalies
Ice	Presence of ice-related mammals Migration of biota (in and out) caused by ice Limited production in winter
Cold bottom water	Outmigration of biota Higher mortalities and lower growth of benthic and demersal biota Accumulation of generations
High runoff	Low salinities (near coasts) High turbidities Presence of eurohaline fauna
Sluggish circulation	Local biological production Local pelagic spawning
<i>Biological features</i>	
High production and slow turnover	High standing stocks
Fewer species (than in lower latitudes)	Few species quantitatively very dominant
Large numbers of marine mammals and birds	High predation by apex predators
Pronounced seasonal migrations	Great local space and time changes of abundance
<i>Fisheries resource features</i>	
Pollock dominant semidemersal species	Flexible feeding and breeding habits, special environmental adaption
Yellowfin sole dominant demersal species	Abundant benthos food supply
Herring and capelin dominant pelagic species	Important forage species in the ecosystem
Abundant crab resources	Large, relatively shallow shelf Few predators on adults, special environmental adaption
Abundant marine mammals	Abundant food supply, no enemies, insignificant hunting. Compete with man for fishery resources
<i>Man-related features</i>	
Fisheries development rather recent	Ecosystem in near-natural state, not yet fully adjusted to effects of extensive fishery
Little-inhabited coasts	Ample space for breeding colonies of mammals and birds Very limited local fisheries, no pollution

March

Air temperatures begin to increase and ice approaches maximum extent except in unusually cold years (e.g., 1956, 1971, 1976), when the maximum

occurs in April. Sea surface temperatures show little change, and the water column under the ice is isothermal at -1.8°C down to depths exceeding 100 m—even deeper (200 m) at the northwest edge of the shelf where ice cover has existed since December.

Figure 36-2. Schematic diagram of March distribution of halibut, herring, pollock, and yellowfin sole. Long dashes indicate probable extension of shelf range; broad line indicates ice edge.)

Cold water (0 to -1.8°C) in the midshelf area has nearly reached a maximum southern extent just north of the central Alaska Peninsula, but the warm-water ($\sim 3^{\circ}\text{C}$) area north of the western edge of the Alaska Peninsula remains.

It is not known whether the progressive cooling of shelf edge and upper slope water (which continues until the ice edge retreats) or spawning behavior triggers a southward movement of various stocks, but increases are apparent in numbers of pollock south of the Pribilof Islands and spawning continues in the area from the 100-m isobath seaward (Fig. 36-2). Why spawning does not occur farther inshore is not known. Halibut and yellowfin sole, present in the spawning area, are about to begin to migrate shoreward. Herring north of the Pribilof Islands begin shoreward migrations to coastal spawning sites; to what extent the ice cover affects these migrations is not known, except that herring have wintered north of Norton Sound under the ice (as in the Gulf of Bothnia and the Gulf of Finland).

April

Air temperatures begin to rise above 0°C immediately north of the Alaska Peninsula but are below freezing throughout the rest of the area. Although in cold years the ice edge may still be advancing slowly southwestward and sections of it may be driven to the coast of the Alaska Peninsula as far west as Unimak Pass (e.g., in 1956), usually a corridor

Figure 36-3. Schematic diagram of April distribution of halibut, herring, pollock, and yellowfin sole. (Solid triangles indicate spawning areas; long dashes indicate probable extension of shelf range; broad line indicates ice edge.)

remains ice-free north of the western half of the peninsula. Ice over the southern half of the shelf reaches maximum thickness: ~ 90 cm inshore near Nunivak and 45 cm near the Pribilof Islands. Isothermal conditions through the water column occur over most of the area as air temperatures north of the peninsula rise above 3°C . In Bristol Bay where air temperatures over land have increased to 5°C , some runoff occurs and bottom temperatures rise from -1.8 to 0°C . However, northward along the coast winter water temperatures of -1.8°C prevail.

Pollock spawning continues seaward to the 100-m isobath with maximum concentrations of fish still north of Unimak Island (Fig. 36-3); although concentrations exist seaward of the shelf the abundance is not known. Herring are proceeding shoreward; the effects of the ice cover and the -1.8°C isothermal condition in the water column at midshelf are not known. In 1976, a cold year, spawning activity along the peninsula seemed to increase, but the apparent increase may have been the result of the intense survey effort that year; in 1979, a warm year, early and extensive spawning occurred in the Togiak area. Yellowfin sole also begin easterly shoreward migrations from north and south of the Pribilof Islands, apparently in advance of halibut.

May

Air temperatures rise rapidly, and below 0°C conditions occur only generally north of St. Law-

rence Island. Retreat of the ice edge is equally dramatic, reaching nearly 10 km/d in the central shelf area. The water shows little evidence yet of a surface layer, but surface temperatures above 4 C occur north of Unimak Island, about 2 C in inner Bristol Bay, and above 0 C northward along the coast to Norton Sound; but runoff is still limited and coastal salinities generally remain above 30‰.

Pollock spawning has reached a peak and adult concentrations shift northward along the outer shelf and slope; planktonic eggs and larvae largely concentrated near the surface are dispersed by currents off and across the shelf but primarily northwest along the shelf edge. Halibut extend into the midshelf area southeast of the Pribilof Islands, their northeasterly advance hindered or at least affected by the midshelf cold zone. Yellowfin sole begin to migrate east along two paths, one north and the other south of the Pribilof Islands (although in May 1978 and 1979, warm years, yellowfin sole were observed feeding on herring spawn in northern Bristol Bay). Herring reach the coast from the Alaska Peninsula to Nunivak Island and spawning occurs in these areas (Fig. 36-4). Juvenile salmon appear along the coast and maturing salmon at the outer shelf. It is interesting to note that at this time the male fur seals, heavy predators particularly on pollock and salmon, have arrived at the Pribilof Islands (see Harry, volume 2 of this book).

June

Air temperatures (offshore) range from 3 to 8 C and the ice edge retreats rapidly (30 km/d) into the Chukchi Sea, except at the western side of the Bering Strait and northern Gulf of Anadyr. Surface water temperatures throughout the area are above 0 C (14 C in Norton Sound, 6 C north of the peninsula and in Bristol Bay); and water is 0-2 C in the midshelf area between the 40 and 150-m isobaths north of the Pribilof Islands to the Gulf of Anadyr. Surface and inshore dilution from ice melt and runoff is first to become evident; coastal surface salinities reach a minimum in Norton Sound (16‰), enhancing surface warming.

Herring spawning occurs along the coast north of Nunivak Island to Norton Sound; to the south spawning is complete and herring disperse along the coast (Fig. 36-5). Pollock move shoreward to the 60- to 70-m isobaths east of the Pribilof Islands and northward to 60°N along the outer edge of the shelf; spawning has largely ceased, larvae have been transported northward along the shelf edge, and juveniles extend inshore along the coast. Halibut extend into inshore areas as they conduct feeding migrations throughout the shelf area. Yellowfin sole continue to have two thrusts shoreward, one north of the Pribilof Islands and the other along the north side of the peninsula and northward in outer Bristol Bay where overwintering juveniles have an opportunity to

Figure 36-4. Schematic diagram of May distribution of halibut, herring, pollock, and yellowfin sole. (Solid triangles indicate spawning areas; long dashes indicate probable extension of shelf range; broad line indicates ice edge.) Some salmon present.

Figure 36-5. Schematic diagram of June distribution of halibut, herring, pollock, and yellowfin sole. (Long dashes indicate probable extension of shelf range; short dashes indicate juvenile distribution, where known or different from adults.) Salmon occur throughout the shelf area.

accompany adults en route to spawning grounds. Salmon occur throughout the southern part of the shelf area; king salmon reach the Yukon River mouth, and the extensive sockeye salmon runs reach Bristol Bay before the end of the month.

July

Air temperatures approach 10 C and sea-surface temperatures of 10-12 C occur in the Gulf of Anadyr, 3-5 C at the west side and 5-7 C at the east side of Bering Strait, 14-16 C in inner Norton Sound, 10-12 C southward along the coast to Unimak Pass, and 5-7 C in midshelf areas seaward over the basin. Bottom temperatures in the midshelf area from the Gulf of Anadyr to the Alaska Peninsula range from -1 to 1 C; in Bering Strait, 1-2 C; in inner Norton Sound, 5-9 C; along the coast from Norton Sound to Unimak Pass, 7-10 C; and along the shelf edge, 2-4 C. Off-shore dilution due to river discharges along the Alaskan coast continues and minimum salinity values occur in outer Bristol Bay (28‰).

Herring spawn along the coast northward of Norton Sound to Kotzebue Sound; no information is available on the Gulf of Anadyr stocks (Fig. 36-6). Yellowfin sole spawning begins northwest and south-east of Nunivak Island, the two stocks apparently remaining separated; however, both spawning areas remain inshore of the mid-shelf cold front in an area of general northerly flow. Pollock distribution

remains the same as the previous month, adults generally seaward of the 50-m isobath but juvenile pollock and herring remaining in inshore areas. Halibut continue to expand their range northward over the shelf. Salmon are ubiquitous, adults entering coastal areas and rivers from Cape Navarin to the Alaska Peninsula to spawn, and smolts becoming acclimated to the oceanic salinities and moving seaward across the shelf to begin the oceanic phase of their life cycle.

August

Air temperatures over the shelf have reached a maximum in July and begin to decline, but are still 5-10 C. However, along the coast sea-surface temperatures continue to increase; maximum values (15 C) occur in Norton Sound and temperatures in excess of 10 C occur along the coast from the Alaska Peninsula to Bering Strait and even in Kotzebue Sound. On the western side of Bering Strait values of 2-4 C occur along the coast but these rise to 7 C along the eastern side of the Gulf of Anadyr and 13 C along the west. Mid-shelf values range from 7 to 9 C, but at the shelf edge they are 9-10 C. There is a marked difference in bottom temperatures, which range from 9 C to -1 C northward to Bering Strait; whereas 10 C values generally occur along the coast from Bering Strait southward to the Alaska Peninsula. Values in mid-shelf range from -1 C in the north to 1 C in the south, and values of 3 C persist along the shelf edge. Maximum stability in the surface layer and maximum thermocline occur in the mid-shelf area. Surface salinities in east Norton Sound have increased to 20‰, whereas dilution in inner Bristol Bay is still evident (26‰).

Yellowfin sole spawning is completed and these stocks disperse over the shelf (Fig. 36-7). Herring begin to migrate seaward off Nunivak and Unimak Islands to wintering grounds; the distribution of halibut and pollock remains largely unchanged. Most of the maturing salmon have reached spawning sites but smolts and juveniles continue moving seaward in undefined paths.

September

Air temperatures drop to 3-8 C and although maximum inshore sea surface temperatures of 10 C prevail from the Alaska Peninsula to Norton Sound, values of 6-8 C occur in the Gulf of Anadyr and 5-8 C in the mid-shelf area. Bottom temperatures remain relatively unchanged from August, as do general salinity conditions.

Of particular interest at this time is the apparent seaward intrusion of warm, dilute, obviously coastal

Figure 36-6. Schematic diagram of July distribution of halibut, herring, pollock, and yellowfin sole. (Solid diamonds indicate spawning areas; long dashes indicate probable extension of shelf range; short dashes indicate juvenile distribution, where known or different from adults.) Salmon occur throughout the shelf area.

Figure 36-7. Schematic diagram of August distribution of halibut, herring, pollock, and yellowfin sole. (Long dashes indicate probable extension of shelf range; short dashes indicate juvenile distribution, where known or different from adults.) Salmon occur throughout the shelf area.

water, southwest of Nunivak Island (see Ingraham and Section I of this book). If not an artifact caused by the averaging of data, this intrusion could have a pronounced effect on the return seaward of herring and yellowfin sole, and possibly other stocks.

The offshore migrations of herring west of Nunivak Island and northwest of Unimak Island continue, and yellowfin sole begin to disperse from the spawning area off Nunivak Island (Fig. 36-8). Although it is apparent that yellowfin sole subsequently range over a large portion of the shelf, it is believed that the major concentrations remain west of Nunivak Island and it is from this location that the two major stocks begin to migrate seaward to wintering grounds at the outer shelf. When juveniles that will winter in coastal areas separate is not clear. Halibut and pollock distributions are apparently unchanged. And salmon stocks, both adult and juvenile, have largely cleared the shelf area by the end of the month, but stragglers may remain.

October

Air temperatures markedly decline and values below zero occur over the northern half of the shelf; ice begins to form in the Gulf of Anadyr, the west side of Bering Strait, and east Norton Sound. Oceanographic data are too fragmentary to indicate mean conditions over the shelf but some features are

apparent. Surface and bottom temperatures of 6-8 C occur along the coast from the Alaska Peninsula to Norton Sound. Mid-shelf surface temperatures of 5-7 C occur and bottom temperatures remain largely unchanged with only minor increases due to downward diffusion of heat but marked changes where there is seaward advection. Coastal salinities increase above 30‰ except in Norton Sound, where values of 27-28‰ are reached.

Herring off Cape Navarin and Unimak Island have mostly reached wintering grounds and the main stock northwest of the Pribilof Islands appears between the 50- and 100-m isobaths south of St. Matthew Island (Fig. 36-9). Some juvenile yellowfin sole enter Bristol Bay where they will winter, but the two main adult stocks move southwestward and southward, returning to winter grounds. Halibut and pollock stocks have begun to retreat to the shelf edge. Why this occurs while even inshore coastal temperatures, particularly on the bottom, are warmer than those they will overwinter at (3-4 C) and no extensive offshore ice occurs is unknown. By the end of the month most of the fur seals will have left the Pribilof Islands and moved southward out of the Bering Sea (see Harry, volume 2 of this book).

November-December

Although essentially only air temperature (-10 to 3 C in November, and -15 to 0 C in December) and

Figure 36-8. Schematic diagram of September distribution of halibut, herring, pollock, and yellowfin sole. (Long dashes indicate probable extension of shelf range; short dashes indicate juvenile distribution, where known or different from adults.) Salmon occur throughout the shelf area.

Figure 36-9. Schematic diagram of October distribution of halibut, herring, pollock, and yellowfin sole. (Long dashes indicate probable extension of shelf range; broad line indicates ice edge.)

ice cover data are available, one can presume that the effects of the wind mixing and subsequent ice cover will quickly dominate any downward diffusion of residual heat in the water column. Both cooling effects will reach depths exceeding 100 m; the former takes place within a day or so, the amount of cooling depending on water temperatures in the surface layer, and the latter requires one to several weeks. However, it is prolonged ice cover that gradually drives bottom temperatures on the shelf down to -1.8°C and the freezing-out of fresh water results in higher salinities. Nevertheless, temperatures of $3-4^{\circ}\text{C}$ still occur on the bottom at the shelf edge and upper slope. The ice edge in November generally extends from Bristol Bay seaward of Nunivak and St. Lawrence Islands into the western Gulf of Anadyr. By December it extends from outer Bristol Bay past St. Matthew Island to Cape Navarin.

Herring and yellowfin sole are believed to reach offshore wintering grounds in November (Fig. 36-10). Herring overwintering in the Norton Sound area under the ice may experience temperatures higher than -1.8°C in the deeper areas if wind mixing did not establish neutral stability in the water column before ice cover was formed. Pollock retreat to the 100-m isobath by the end of December; at this time halibut have largely retreated seaward of the 200-m isobath.

RESOURCE ENVIRONMENT RELATIONS

The life histories of the dominant fish (except for halibut) selected for discussion in this section are quite diverse. Harden Jones (1977) suggests that fish movements can be summarized in the form of triangular patterns in which adult fish move between their spawning, feeding, and wintering areas in a sequence which depends on the season in which they spawn: spring spawners (here, pollock and herring) follow a clockwise sequence of feeding-wintering-spawning, and autumn spawners (no representatives among the species studied here) a counterclockwise sequence of feeding-spawning-wintering. Such a scheme does not account for halibut, which spawn in winter in the wintering area, or yellowfin sole, which spawn in the summer far from the wintering area.

Plausible ranges and time scales of advection (passive transport) and migrations are shown schematically in Figure 36-11. The average advection (~ 1 km/hr) and average migration (~ 3 km/hr) are indicated, relating time to distance; periodic migrations such as semidiurnal and seasonal migration are also shown. The expected relative magnitudes of changes of biomass of a given species as a result of different periodic migrations as well as possible long-period "drift" within the system (which could result in long-period changes in abundance depending on transgressions of pelagic fish on and off the shelf and demersal fish into and out of the areas along the shelf) are shown in Fig. 36-12. Migration speeds are

Figure 36-10. Schematic diagram of November-December distribution of halibut, herring, pollock, and yellowfin sole. (Long dashes indicate probable extension of shelf range; broad line indicates ice edge.)

not known, except perhaps for Bristol Bay sockeye salmon, but some estimates of magnitudes of shelf distances traversed and required speeds for the eastern Bering Sea have been compiled (Table 36-2). Data for similar species in the North Sea indicate that these estimated speeds are considerably lower than might be expected and Bering Sea stocks may have considerably more mobility over the shelf than heretofore believed.

Although environmental conditions are variable and stock distributions fluctuate, there are apparently no environmentally related indices, other than generally defined warm or cold years, that can be used to forecast successful or unsuccessful spawning or year-class strengths, anomalous distributions or migration paths, or fluctuations in annual abundances. Even though in the foregoing summaries of monthly events only mean conditions were considered, and events could be advanced or delayed as much as a month as a result of warm or cold conditions, the initial primary considerations for determining distributions were largely depth and season: for example, halibut occur in winter seaward of the 200-m isobath; adult pollock are generally seaward of 100 m in winter regardless of the position of the ice edge and seaward of 50 m in summer; juvenile pollock in summer are found shoreward of the 100-m isobath; the yellowfin sole wintering area is largely between the 100- and 200-m isobaths but juveniles overwinter in inner Bristol Bay under the ice, where the shallow depths assure that the water temperature will be -1.8°C ; herring overwinter largely between the 100- and 200-m isobaths again apparently without

Figure 36-11. Ranges of advection and migration of stocks in marine ecosystems.

Figure 36-12. Relative changes in abundance of stocks due to advection and migration.

any regard to the ice edge and some inshore stocks overwinter in ice-covered coastal regime. In only a few instances is there specific evidence of environmental influences. Yet if one approaches these conditions or events from a fishery oceanographer's or environmentalist's point of view rather than from that of a management biologist, another perspective is evident.

TABLE 36-2

Speeds of migrating fish

Species	Speed (km/d)
Sole ¹	7-16
Plaice ¹	1-7
Herring ¹	4-30
Salmon (sockeye) ²	54
Salmon (chum) ²	48
Halibut ³	6
Herring ⁴	25
Yellowfin sole ⁴	3-7

¹ Harden Jones (1977)—North Sea area

² Kondo et al. (1965)—E. Bering Sea shelf

³ Noviko (1970)—E. Bering Sea shelf

⁴ Estimated from spawning migration—E. Bering Sea shelf

Winter

Most discussions of productivity deal with conditions on the roughly one million km² of shelf, yet it is obvious that most of the biomass (excluding mammals) is restricted to a small fraction (less than a fifth) of the total shelf area for nearly half the year, November to April. Although one might argue that feeding in winter is usually drastically reduced, it should be obvious that bottom temperatures at the

shelf edge (3-4 C) during the winter are several degrees higher than in the mid-shelf area throughout the summer (-1 to 2 C), even though summer inshore temperatures may reach 10-15 C. Predator and prey are crowded together and there is evidence at other times and places that halibut consume yellowfin sole; pollock are primarily cannibalistic (Takahashi and Yamaguchi 1972). (1) What effects do changes in the continuity and extent of the anomalous 3-4 C bottom temperatures on the outer shelf and upper slope have on species and biomass distributions, and to what extent does predation and growth occur during winter in this narrow zone?

Halibut spawn in January at depth; environmental conditions in the southeastern corner of the Bering Sea basin can be considered relatively constant. The prolonged period from egg release to the time when food is no longer provided by the yolk sac eliminates the availability of forage as a critical factor until early spring. (2) Does the success of a halibut year-class in this area depend on a predominance of onshelf flow along the shelf edge, possibly entrainment of larvae in local eddies, or is this stock maintained by transport of eggs and larvae from the North Pacific Ocean through Aleutian Island passes, with spawning in this area providing the basis for stocks on the Asian (Cape Navarin/Cape Olyutorski) coast? Of course for management purposes one can wait until individual year-classes are recruited into the fishery, but unless this question is resolved we will not be able to assess the effect of oil or oil spills on halibut (and other groundfish with similar behavior) recruitment in specific areas.

Herring are shown to have possibly three offshore wintering grounds with equivalent bottom topographies, but these represent three areas with distinct environmental conditions. The stock of the area south of Cape Navarin spends a large part of the winter, if not all of it, under the ice at -1.8 C; the major wintering ground north of the Pribilof Islands may have ice cover only in March or April, insufficient time to establish -1.8 C conditions, so that temperatures of 0-2 C prevail; the possible winter concentration between Unimak Island and the Pribilof Islands will encounter temperatures of 2-4 C and experience ice cover and colder conditions only in extremely cold years. (3) Are the three different winter temperature regimes critical to the isolation of these three stocks and if so, do the stocks have broader distributions than indicated and do the locations of the stocks change with environmental conditions? Attempts by NWAFC to sample herring in the major wintering ground in February 1978 failed (winter 1978 can be classified as warmer than normal).

Spring

The herring are the first to leave the area of concentrated biomass in winter, and spawning migrations begin in March. This may be an instinctive movement, because ice cover is relatively constant, at depths in excess of 100 m temperature regimes do not change (-1.8 C, 0-2 C, and 2-4 C), and relative darkness at depth does not provide any signal or impetus. The primary factors here must be increasing light evident in upper layers and the absence of ice. (4) Since the extent of mean ice cover usually reaches a maximum in April, are shoreward migrations from wintering grounds near the shelf edge that begin in March conducted under the ice, or do herring migrate southward (or await retreat of the ice edge) and proceed to the coast in open water? If the former were true, herring would arrive in spawning areas on schedule but in cold years they would have to wait until the coast was free of ice to spawn. Although rafting can occur, the ice edge is usually composed of soft ice (vessels have easily penetrated 50-100 km) and such observations could be made; spring distributions and shoreward migration routes could then be ascertained by analyses of satellite imagery of ice cover.

Certainly pollock spawning in spring is a dominant event, although at this time it is difficult to compare its effects to those of sculpin spawning. The abundance of adults increases in spring north of Unimak Island (Low and Akada 1978). (5) Does the progressive cooling of shelf edge and upper slope water north of the Pribilof Islands into April as a result of prolonged ice cover cause a southward retreat of adult pollock or is this a spawning migration?

As might be expected of a successful species, spawning is protracted. Waldron (1978) found that pollock larvae were present near the shelf edge north of Unimak Island in March; if development occurred locally, spawning must have begun in February. However, the larvae were found in a northward-protruding tongue of temperature maxima (4 C) and it is possible that they were advected into the area from south of the Alaska Peninsula, where conditions can be much warmer at this time of the year. Unlike halibut, pollock eggs and larvae occur in the surface layer and are exposed to the vicissitudes of weather from March to July. In spring 1976, eggs were found within the ice edge when an unusual southward thrust of ice occurred, and violent spring storms are the general rule rather than the exception. (6) What are the effects on pollock year-classes of stormy vs. calm weather, ice vs. no ice, and early (February-March) vs. late (May-June) spawning? A simple device that would not only assist in determining

answers but also permit prompt verification and evaluation of conditions would be a moored biological buoy that would sample organisms in the water column (5 m to bottom) at weekly (or shorter) time intervals from February to late spring, when the data would be recovered and evaluated.

Furthermore, pollock spawning appears to be largely restricted to the area seaward of the 100-m isobath, thus very near the Pribilof Islands and the million or so fur seal predators. (7) What conditions, environmental or otherwise, appear to confine adult pollock seaward of the 100-m isobath before, during, and after spawning?

It is not clear whether halibut or yellowfin sole are first to begin shoreward migrations, but the latter have a spawning incentive. (8) Do halibut follow yellowfin sole shoreward to be sure of a source of food or is their shoreward migration and subsequent consumption of benthic organisms simply a seasonal feeding behavior pattern?

In respect to both inshore migrations Harden Jones's (1977) selective tidal transport hypothesis may be applicable. His studies indicate that plaice in European waters are able, because of their shape, which provides lift forces allowing them to glide (or sailplane), to move into midwater, where they spend half their time, taking advantage of tidal flow in the direction of purposeful movement, but returning to the bottom at the turn of the tide. Such a behavior in the stock between the Unimak and Pribilof Islands would permit considerable enhancement of migrations; current data (Dodimead et al. 1963; see Chapter 5, this volume) and tidal model studies at NWAFC (Hastings 1975) indicate a general northeast-southwest oscillation north of the Alaska Peninsula, particularly in Bristol Bay, that would aid migrations from the shelf edge into Bristol Bay and a change in oscillation that would aid a subsequent general northward movement over the shelf. The benefit from such tidal sequences also would aid return movement to the southwestern part of the shelf in fall. (9) Does selective tidal transport give halibut, yellowfin sole, and other flatfish considerably more mobility and migration speed than now believed (~3-6 km/d)?

The onshore delay or aggregation of halibut and yellowfin sole at the cold front north of the Alaska Peninsula is a convincing example of resource-environment relations. But the northward movement of halibut proposed by Natarov and Novikov (1970) and the eastward movement of the yellowfin sole stock north of the Pribilof Islands indicated here raise some doubts as to the absolute validity of this phenomenon; U.S.S.R. studies in the Grand Banks off the east coast have indicated that hunger transcends

discomfort and fish will at times intrude into areas having environmental conditions normally avoided. (10) What are the roles of selective tidal transport and the ability of halibut and yellowfin sole not only to move quickly but to rise easily into the water column in their penetration of the cold, midshelf zone? Studies of benthos and possibly relationships to fish in this area are the subject of section XII of this book.

In regard to salmon movements, over 40 years ago Moulton (1939), in the foreword to a symposium concerning the migrations of salmon, noted that in the problem of accounting for the migration of salmon one enters a field in which science and romance appear to meet. In spite of environmental and tagging studies, the story of what takes place in the apparently supernatural migrations, wherein mature fish return after several years at sea to precisely the stream of their birth, was far from being complete, and explanations of such migrations were all equally open to question. Although considerable information on oceanic distributions and migrations has come from studies conducted since 1953 under the aegis of the INPFC, how salmon find their way in the ocean remains a major and interesting fisheries oceanography problem. Explanations of coastal migrations have the same inadequacies as those of ocean migrations. Less than a decade ago at another salmon symposium Larkin (1975), after reviewing information on how salmon navigate, postulated that they might have a system of collectivizing judgment that could provide a school with remarkable navigational accuracy; if they could detect their confreres by scent, adults could school effectively in the ocean as well as detect coastal regimes carrying the scent of juveniles that had commenced seaward migrations. Although such ideas are perhaps more intuitive than scientific, there are numerous examples of remarkable salmon behavior other than the fact that mature individuals return to natal streams annually and arrive at the coast with consistent and appropriate timing. Such punctuality might be considered supernatural if it were not for their mobility; for example, king salmon migrate over 2,400 km up the Yukon River in 3.5 weeks, advancing at an average speed of 110 cm/sec or about 85 km/d in spite of opposing river flow (Gilbert and O'Malley 1922). Apart from any potential to swim shorter distances at higher speeds, they could transit the shelf area in much less than a week although shelf migration speeds of only 50 km/d are reported (see Table 36-2). Although knowledge of the nature of such movements is scant, there is an apparent order to the sequence of arrivals at river mouths of all species.

Large numbers of king salmon in trawl catches in winter 1979 near Zemchug Canyon came as a surprise to biologists who have studied salmon migrations for decades. Since this canyon is believed to have been cut across the shelf during periods of lower sea level (land bridge) by the Yukon River, if these salmon can be shown to be of Yukon River origin one is inclined to believe that salmon homing instincts have strong environmental imprints. But it is apparent that movements of Yukon River, Norton Sound, and Gulf of Anadyr salmon and those that eventually pass through Bering Strait are largely unknown. Questions concerning the movement of salmon in the eastern Bering Sea in relation to environmental factors are too numerous to list but obviously little is known about movements northward of the Pribilof Islands. (11) What are the migration routes over the shelf of adult and juvenile salmon from Gulf of Anadyr, Arctic Ocean, Norton Sound, and Yukon River areas? Obviously extensive environmental studies must accompany any distributional or tagging problems.

There is much evidence that Pacific salmon respond to environmental conditions (Favorite et al. 1977) and it is apparent that migrations of various species and stocks through the various OCS oil development lease areas (e.g., St. George Basin, Navarin Basin, Norton Sound) cannot be ascertained or predicted until more information is obtained on effects of environmental conditions on shoreward migrating adults and seaward migrating smolts (juveniles) in this area. Oil contamination of river mouths could affect upstream migrations of sockeye, coho, and king salmon and coastal spawning of chum and pink salmon; and spills at offshore sites could affect migrations of salmon of U.S. and U.S.S.R. origin (spills in spring and summer could also be locally harmful to herring, because spawning beds are intertidal).

Summer

It has been pointed out that adult pollock are cannibalistic, preying extensively on juveniles; in summer, juveniles occur largely within the area between the 100-m isobath and the coast, whereas the adults are largely confined seaward of 100 m. (12) Is the presence of juveniles in inshore areas where warm conditions occur a response to environmental conditions including forage or are they found there because it is the only area where predation by adults does not occur; what confines adult pollock seaward of the mid-shelf area?

Herring spawning occurs in northernmost areas in summer, reflecting a northerly sequence with time and perhaps a direct relation to temperature and ice,

and also possibly to the effects of coastal dilution, which is also delayed from that in more southern areas. (13) Can the time of herring spawning at various points along the coast, which can vary from a week to a month, be defined by temperature or salinity conditions, and do advanced or delayed spawning periods affect the success of year-classes, or contribute to reduced or excessive predation on spawners?

The spawning of yellowfin sole in the area off Nunivak Island is an important summer event. It occurs inside one of the sharpest frontal zones on the shelf, between the mid-shelf and coastal sub-domains. Temperatures in the frontal zone at this time increase sharply shoreward, from -1 to 10 C. Thus, spawning occurs in the warmest area of the eastern part of the shelf that lacks the excessive dilution and high turbidity produced by local rivers (unlike inner Bristol Bay and Norton Sound), in an area that has a broad extent of shallow depths (unlike the northern coast of the Alaska Peninsula). The length of time required for larvae to attain mobility or drop to the sea floor is not known, but a northerly flow in this area of 25 cm/sec (~ 0.5 kn) would transport them through Bering Strait in about 20 days unless they were caught in eddies in Norton Sound. (14) What mechanism prevents the dispersal of yellowfin sole larvae into the Arctic Ocean?

It is apparent in the monthly mean distributions of properties for September that there is a marked seaward intrusion of coastal water in the Nunivak Island area that could have a major effect not only on circulation, but on movements of juvenile pollock and herring and the larvae of yellowfin sole. (15) Is this trans-shelf intrusion real or an artifact of averaging data and what is its effect on conditions and events in this area?

Fall

Although little is known concerning precise fall distributions of fishes, it is generally believed that most begin to retreat southward and seaward to the shelf edge by October. Although air temperatures are lower, mean sea surface temperatures have not decreased substantially. Water temperatures in coastal areas have lowered a few degrees but are still high (~ 8 C); and ice appears only in eastern Norton Sound and along the shore. Thus there is little evidence of drastic environmental changes other than in the duration and intensity of light, until November. (16) What triggers fall seaward movements of fish to the shelf edge and what are the nature and relative timing of movements of individual stocks?

CONCLUSIONS

The eastern Bering Sea ecosystem encompasses a vast renewable resource of great potential and value that should be carefully protected and developed but of which we have very little basic understanding. The NWAFC and the North Pacific Fisheries Council are expending great efforts to estimate the various individual resources for management purposes, but there should be an equal effort to understand conditions and processes as well as periodic and aperiodic events and their causes and effects. This can come about only through cooperative, interdisciplinary research. Basic research and resource assessment should continue, but fisheries oceanography studies should combine both of these and provide an integrative focus that is badly needed as exploitation of individual resources of commercial value accelerates.

Nearly a decade ago, when congressional support for fisheries studies was wanting, one argument put forth to justify limited support for marine research was that marine fisheries were a common property resource, could not be protected, and posed real problems in regard to development. The ideal example given was a plot of trees that could be clearly defined spatially and contained a crop that could be protected (from man and diseases), selectively harvested on demand, and selectively reseeded to provide a continuing resource. The eastern Bering Sea shelf, relatively isolated from the broad sweep of ocean currents, is a unique place to consider development of a major marine habitat that can be protected and developed when we have a basic understanding of requirements of individual species and stocks, interactions among them, and their place in the ecosystem. The economic justification is obvious; the value of fisheries products extracted in 1978 was roughly half a billion dollars. The problem of protection is answered by the recent FCMA.

Even from the limited discussion presented, the merit in multispecies and ecosystem approaches to investigating and understanding conditions and processes is readily apparent; such a focus should be considered in all future studies. Because this requires extensive accountability of species interactions and environmental effects, we are at a loss to handle the multiplicity of factors involved without formulating some sort of modeling or simulation techniques. A great deal of preliminary work has been accomplished along these lines at NWAFC and a summary of an existing simulation model (DYNUMES) that presents the quantifications of biomasses omitted in this chapter is presented in the next. Such studies also permit independent assessments of movements

in relation to environmental conditions such as temperature and forage requirements.

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Ecosystem Dynamics in the Eastern Bering Sea

37

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ABSTRACT

The spatial and temporal dynamic aspects of the marine ecosystem in the eastern Bering Sea were investigated with the ecosystem model DYNUMES III. Equilibrium biomasses of major species and ecological groups are presented together with their annual consumption. The latter represents a major portion of the natural mortality, which removes a major part of the annual production of biomass, especially of forage species and faster-growing juveniles of other (larger) species. Examples of the dynamics of the biomasses of some species, with certain restraints on recruitment variations, over a four-year period are also given.

Marine mammals, as apex predators, remove a considerable part of the fish production and the effect of their predation on the marine ecosystem is in many respects similar to that of the fishery.

Some validation of the results of simulation model computations is provided by comparisons with survey results, adjusted by catchability factors.

INTRODUCTION

In the past it has been customary to consider the marine ecosystem and the living resources in it as a quasi-steady-state system. This conclusion has been reached by averaging the quantity of resources over large regions, such as the eastern Bering Sea, and over long time units, such as a year or more. However, conditions in marine ecosystems are not stable in space and time, but are affected by various seasonal changes (e.g., feeding and spawning migrations) and by environmental anomalies (e.g., varying extent of ice cover), all of which are rather pronounced in the Bering Sea.

The objectives of this study were to investigate the effects and spatial and temporal magnitudes of the dynamics of the Bering Sea ecosystem, including the source and sink areas of biomasses of various species and the seasonal changes of these sources and sinks as well as the changing predator-prey relations. The eastern Bering Sea contains large numbers of marine mammals as apex predators whose numbers vary seasonally. Their effects on the ecosystem were also

investigated and are briefly reported. Supporting data and details are being reported currently in various NWAFC Processed Reports.

MAJOR PROCESSES IN THE MARINE ECOSYSTEM AND A REVIEW OF THE ECOSYSTEM SIMULATION MODEL DYNUMES III

The major dynamic processes in any marine ecosystem which affect the abundance and distribution of species are listed in Fig. 37-1 together with the major factors affecting these processes and their results.

The most difficult to reproduce quantitatively are the results relating to the spawning process, i.e., the larval recruitment and the early mortalities of larvae. After many decades of intensive research on stock-recruitment problems, we now know that no simple relation exists between the size of the spawning stock and the amount of recruitment to an exploitable stock. It can only be stated in general terms that proportionally larger recruitment can result from a small spawning stock and proportionally smaller recruitment can result from a large spawning stock. The numerous hypotheses on proper food availability to larvae and subsequent effects on recruitment size have not been proved either. Many fisheries scientists are inclined to believe now that recruitment numbers to exploitable stocks are controlled mostly by predation in larval and juvenile stages (Hempel 1978, Rothschild and Forney 1979) and that recruitment might be largely influenced by environmental anomalies (Garrod and Colebrook 1978).

The conditions affecting growth and reduction of species biomasses are generally known, although refinement of knowledge on the small variations in the growth of individual species is continuing.

Figure 37-1. Major dynamic processes in the marine ecosystem.

Predation has been included traditionally in the all-encompassing term "natural mortality," of which predation constitutes the major portion in most species. In order to remedy this shortcoming the predation mortality is computed within ecosystem models in great detail, using data on food requirements for growth and maintenance and space- and time-variable food composition. Very few fish stomach analysis studies have been accomplished on Bering Sea stocks. Those available usually list only the frequency of

occurrence of food items in the stomach, making these studies of little quantitative value. Fortunately recent European studies of fish feeding habits (e.g., Daan 1973) have established that feeding in the marine environment is largely dependent on the size of the fish and the size and availability of the food; this knowledge has been incorporated in our DYNUMES (Dynamical Numerical Marine Ecosystem) model. In addition, special studies at NWAFC have quantitatively ascertained the age-

dependent spawning stress and senescent mortalities (Granfeldt 1979a), and these also have been included in the DYNUMES model.

Because of spatial changes in factors affecting growth and in predator and prey distributions, considerable spatial and temporal differences in the growth and decline of biomasses occur (these are described later). All of these interactions are non-linear and therefore can be evaluated quantitatively only with three- or four-dimensional ecosystem models such as DYNUMES. Few epibenthic organisms are sessile during most of their lives. Most species undertake migrations for various reasons; seasonal migrations are connected most often with seasonal changes in the environment, both in relation to the availability of food and to the search for optimum environmental conditions (e.g., temperature). In addition, pelagic organisms are advected with seasonally changing currents. Life-cycle dependent migrations are for spawning, feeding, and predation avoidance. Very little is known of the migrations of fish and especially of the distribution of juveniles in the eastern Bering Sea. Only the seasonal migrations of some flatfishes are approximately known in the North Pacific (Alverson 1960).

The static conditions usually implied in past studies of ecosystem productivity and of the effects of fishing will not give quantitative answers when prey is quasi-stationary and predators migrate or vice versa. Some of these migration effects are shown schematically in Fig. 37-2. Consider that there is a given benthos biomass as a fish food resource at the section a-b. Under stationary conditions this benthos biomass is grazed by the stationary predator biomass (a-b) at this location. However, if the predator moves with a speed C through this section, the "upmigration" of biomass A can prey on the same standing stock of benthos at the section a-b during migration. If the speed is doubled, the biomass A+B can prey on the same benthos biomass (time factor must also be considered above). This applies also to fishing: in the stationary case, the fish is caught (and sampled) as representing the biomass present at all times. However, with varying migration speeds and quantities of biomass passing through the section, the effect of a constant fishery for different segments of the population would be different. This difference becomes more complex if the biomass age composition also varies with time as the fish biomass moves through section a-b. The second part of the figure shows the self-explanatory effect of predation caused by separation and/or overlap of predator and prey. These concepts are simulated numerically in the model.

Figure 37-2. Schematic presentation of two effects of migration. (I) With a given migration speed c the biomass A passes through the section a-b. Doubling either the speed or time, biomasses A + B pass through the same section. (II) Schematic example of the migration of a predator into the region of a given prey.

The extensive and relatively complex ecosystem simulation model DYNUMES has been described elsewhere (Laevastu and Favorite 1978a, and in press). All necessary species-specific data, such as growth rate, temperature preference, and food item preference, are obtained from available empirical data and introduced into the model. The initial distributions of groups of species are prescribed at each grid point in a geographically oriented grid (Fig. 37-3A), together with other necessary space- and time-dependent data such as depth, monthly surface and bottom temperatures, ice cover, and migration speed (U and V components).

The numerous computations are accomplished at each grid point at each time step (e.g., monthly). There are four basic formulas and a great number of auxiliary computations. The first basic formula is the biomass balance formula:

$$B_{i,t,n,m} = B_{i,t-1,n,m} (2 - e^{-g_{i,t,n,m}}) e^{-m} - C_{i,t-1,n,m} \quad (1)$$

where $B_{i,t-1,n,m}$ is the biomass of species i in previous time step ($t-1$) at the location n,m ; g is the growth coefficient which is a function of temperature, season, and availability of food; m is a mortality coefficient (senescent mortality and spawning stress mor-

Figure 37-3. Computation grid of DYNAMES for the eastern Bering Sea (A) and the fisheries management and statistical areas (B).

tality), and C is predation (mortality) as computed in previous time step:

$$C_{i,t,n,m} = \sum_j C_{i,j,n,m} \quad \text{and} \quad (2)$$

$$C_{ij} = R_{j,t,n,m} \Pi_{j,i,t,n,m} \quad (3)$$

where $R_{j,t,n,m}$ is the total food requirement of species j and $\Pi_{j,i,t,n,m}$ is the fraction of food of species j which consists of species i . This fraction varies in space and time (t,n,m) depending on the availability of suitable food items.

The migrations are computed with an "upmigration" interpolation finite difference formula:

$$B_{t,n,m} = B_{t-1,n,m} - (t_d |U_{t,n,m}| UT_{n,m}) - (t_d |V_{t,n,m}| VT_{n,m}) \quad (4)$$

where: t_d is the length of time step, U and V are migration speed components, and UT and VT are migration gradients of biomass.

EQUILIBRIUM BIOMASSES IN THE EASTERN BERING SEA AND THEIR LONG-TERM DYNAMICS

If the growth of all individual biomasses in a given ecosystem equals their removal by predation, fishery, and other causes of mortality, we can talk about equilibrium biomasses which can be sustained in this given ecosystem. These equilibrium biomasses can be determined under certain conditions: by ignoring migration effects and using the Bulk Biomass Model for given areas (Laevastu and Favorite 1978b), and by finding a unique solution for equations (1) to (3), provided part of the consumption (C) is predetermined.

The predetermined consumption for the eastern Bering Sea has been obtained by computing the consumption by marine mammals as the first step in the computations, using prescribed monthly numbers of mammals and keeping their food requirements and food consumption constant. Despite a number of other minor limitations of the Bulk Biomass Model (e.g., the food composition can vary only to a limited extent), the error of the results of this model does not exceed ± 30 percent of the given value of the biomass. The equilibrium biomasses in NAWFC Statistical and Management Areas 1, 2, and 3 (Fig. 37-3B) are given in Table 37-1. These biomasses are the "minimum sustainable equilibrium biomasses" (Laevastu and Favorite 1978a) obtained by using the highest plausible growth rates and the lowest plausible food requirements. The equilibrium biomasses (Fig. 37-4) are grouped by three different regimes (pelagic, semidemersal, and demersal). The semidemersal species dominate all other ecological groups (12 million mt), mainly because of their more flexible feeding habits. The biomasses of pelagic and demersal species are about equal (ca. 7.5 million mt each). The most abundant species is pollock (ca. 9 million mt), followed by cottids and other smaller, noncommercial demersal species (4 million mt), and capelin, other smelts, and sand lance (3.5 million mt). Salmon, which occur seasonally, are not included in Table 37-1 and Fig. 37-4.

The biomasses of demersal stocks decrease more than one order of magnitude from Area 1 to the deep ocean regime, Area 3, whereas there is little relative change in the biomasses of the pelagic species between the continental shelf and deep ocean regimes. The biomasses of semidemersal stocks also decrease from the continental shelf regime to the deep ocean regime, mainly because of the disappearance of the benthic food resource. The semidemersal species live a pelagic life and consume pelagic food over the deep water.

Figure 37-4. Equilibrium biomasses of three different regimes in the eastern Bering Sea.

Very little information has been available about the benthos. However, recent work, much of which is reported in Section XII of this book, improves this situation considerably. The total equilibrium biomasses require about 50 g/m² standing stock of benthos. The existence of this standing stock is entirely possible if we compare the eastern Bering Sea with the well-investigated Barents Sea.

Quantitative zooplankton data from the eastern Bering Sea have been sparse; recent information is given in Section X of this book. Soviet studies in the early 1960's were quantitatively deficient, giving only the minimum standing stocks of copepods and no quantitative data on abundant euphausiids (Laevastu et al. 1976). Since the total equilibrium biomasses consume about 50 g/m² of zooplankton, the annual production of zooplankton must be at least this amount.

The annual turnover rate (last column in Table 37-1) (turnover rate = annual consumption/mean standing stock) provides information on the predation

and other mortalities of the species or groups of species. In the marine ecosystem, the younger, smaller organisms are most vulnerable to predation. The growth rate of the biomass is highest at the younger ages. Thus the growth rates determine the length of the period during which a given species is most vulnerable to predation. The distribution of biomass with age and the predation-vulnerability (Fig. 37-5) are shown for two species as an example.

The long-term dynamics of the biomasses in the marine ecosystem can be studied with the DYNUMES and BBM models after determining the equilibrium biomasses by introducing a cause of change in behavior or abundance of any species in the ecosystem. The results of such studies have limited reliability beyond a few years because of the uncertainty in predicting the survival to recruitment (spawning success). An example of the change in biomasses of a few species in Area 1 (see Fig. 37-3B) over four years, assuming that larval recruitment was directly proportional to the spawning biomass present, rather than a

Figure 37-5. Mean biomass and its annual production distribution with age in pollock and yellowfin sole. The portion of biomass highly vulnerable to predation is indicated.

TABLE 37-1

Equilibrium biomasses and ecosystem internal consumption of some species/groups of species in the eastern Bering Sea (in 1,000 tons)

Species or groups of species	Total biomass, in 1,000 tons				Exploitable biomass	%	Consumption, in 1,000 tons			Annual turnover rate
	Area 1	Area 2	Area 3	Eastern Bering Sea			Area 1	Area 2	Area 3	Eastern Bering Sea
Halibut, turbot Flathead sole, arrowtooth flounder	260.5	142.1	8.8	411.3	54		66.9	31.7	2.8	101.4
Yellowfin sole, rock sole	945.1	236.4	17.0	698.4	54		147.7	69.6	7.0	224.3
Other flatfish	661.8	444.7	24.0	1,130.5	45		174.1	87.6	9.2	270.9
	492.3	363.5	26.5	882.3	28		220.2	111.2	14.3	345.7
Cottids	2,478.1	1,450.0	189.5	4,117.6			1,347.2	664.7	115.8	2,127.7
Cods	593.9	391.5	50.6	1,036.0	72		301.2	168.6	31.5	501.3
Sablefish	71.3	37.7	18.0	127.0	40		18.6	10.2	1.4	30.2
Pollock	5,513.4	2,997.3	702.5	9,213.2	70		2,390.9	1,135.2	464.2	3,990.3
Rockfishes	1,009.9	477.3	139.0	1,626.3	30		445.2	210.9	65.8	721.9
Herring	1,219.1	578.5	171.4	1,968.9	30		832.9	399.1	125.6	1,357.6
Capelin, sand lance	2,165.8	1,047.8	292.3	3,505.8			1,550.7	749.5	219.2	2,519.4
Mackerel	701.3	330.4	131.9	1,163.6	45		387.9	183.0	78.3	649.2
Squid	523.3	248.4	498.3	1,270.1			621.5	307.5	668.1	1,597.1
Crab	522.2	309.9	14.3	846.4	40		180.0	93.1	5.9	279.0
Shrimp	427.3	479.9	24.3	931.5	65		350.0	187.8	14.2	552.0
Predatory benthos	466.6	378.0	41.7	886.3			198.0	138.4	20.8	357.2
Infafauna	13,401.3	10,517.4	443.3	24,362.0			13,813.0	9,085.9	546.4	23,445.3
Epifauna	10,433.3	7,006.9	452.0	17,892.0			6,471.8	3,845.4	305.0	10,622.2

function of the relation between equilibrium biomass and actual biomass (i.e., as conventionally assumed that large spawning biomasses produce proportionally smaller recruitment and vice versa) is shown in Fig. 37-6. The interactions of the biomass fluctuations are rather complex and would take too much space to analyze here, but computer printouts are available for this study. It should, however, be pointed out that there are "natural," quasiperiodic fluctuations of biomasses of considerable magnitude in the marine ecosystem. The period can be from a few years to more than a few decades and the magnitude can be considerable (e.g., the biomass can be a fraction of a few tenths to several times its long-term mean value). These fluctuations can have manifold causes in addition to man (fishing). They can, to a certain extent, be studied with ecosystem models.

SEASONAL DYNAMICS OF THE BERING SEA ECOSYSTEM

There are two basically different causes of seasonal spatial and temporal changes in the biomasses. The changes in abundance are caused by seasonally changing growth, predation (and other causes of mortality), and production and release of eggs and milt; the changes in distribution are caused by seasonal migrations of species and environmental interactions. The results of both major causes of seasonal

Figure 37-6. "Natural" changes of biomasses of yellowfin sole, Atka mackerel, pollock, and herring in the fisheries management Area 1 in the eastern Bering Sea.

dynamics must be viewed spatially. Unfortunately, little consideration has been given to the spatial aspects of biomass (and ecosystem) dynamics in the past, mainly because of difficulties in empirical study by nonsynoptic resource surveys. However, the gridded ecosystem models with spatial resolution such as DYNUMES make such studies possible.

Examples of spatial and temporal aspects of biomass dynamics (Figs. 37-7 and 37-8) depict the biomass sources and sinks in February and August of two different-sized groups of pollock (juveniles, <22

Figure 37-7. Sources and sinks of juvenile pollock (<22 cm long) in February and in August in the eastern Bering Sea (in 100 kg/km²).

Figure 37-8. Sources and sinks of maturing pollock (22-45 cm long) in February and in August in the eastern Bering Sea (in 100 kg/km^2).

cm, and maturing pollock, 22 to 45 cm). *Source* refers to the area where biomass growth in a given time interval (month) exceeds its losses by predation, fishery, and other causes of mortality; *sink* refers to the opposite condition, i.e., losses exceed growth. The sources and sinks of older pollock (>45 cm) are not shown because this size group has only sinks (large senescence and spawning mortality, fishery, and small growth rate) which are roughly proportional to biomass present.

The sources and sinks of all species change because of spatial and temporal changes of the processes which cause them. There is usually a sink at the periphery of the distribution of the biomass. This sink is usually compensated for by outmigration from the center of main distribution (spreading). There is a nearly continuous source of pollock off the continental slope over the deep water; during winter this source area is displaced southwest where the temperature of the water is higher, allowing higher growth rates.

The distribution of the three different age groups of pollock in August is shown in Figs. 37-9, 37-10, and 37-11. A partial separation of juvenile and old pollock is brought about by cannibalistic predation of older pollock on its own juveniles. The highest concentration of biomass of older pollock is found

off the continental slope, whereas the juveniles are found mainly on the continental shelf.

The effects of seasonal depth-migrations of yellowfin sole on its distribution changes are shown in Figs. 37-12, 37-13, and 37-14. The seasonal depth migrations of flatfish were investigated by Alverson (1960) and on the basis of his work it was assumed that yellowfin sole migrate from deep water into shallow water during May and June (Fig. 37-12) and back into deep water in October and November (Fig. 37-13). A migration speed of 3 km/day was assumed in the model. It is not always possible to determine a single cause of seasonal migrations, which can be spawning, search for food, or search for acceptable or optimum environmental conditions.

This resulting monthly change of biomass can be considerable in an area where active and extensive migrations occur (see Figs. 37-12 and 37-13). These seasonal migrations have profound effects on other biota as well as on the evaluation of fishery resources by trawling surveys. For example, flatfish are dependent on benthos as a food source and migrations cause heavy grazing of benthos in some areas during some seasons, allowing a recovery period during other seasons. A proper trawling survey evaluation must account for seasonal migration to avoid erroneous results.

Figure 37-9. Distribution of juvenile pollock (<22 cm long) in August in the eastern Bering Sea (tons/km²).

Figure 37-10. Distribution of "maturing pollock" (22-45 cm long) in August in the eastern Bering Sea (tons/km²).

Seasonal migrations are affected by various environmental anomalies; water temperature anomalies, especially at high latitudes as in the Bering Sea, are usually the most pronounced and the easiest to observe. Furthermore, more is known of the effects of temperature on the species than of the effects of other environmental variables. Two dynamic effects of temperature anomalies are included in the DYNUMES model: the "forced" migration of most species out of areas with subzero bottom temperatures (including a slightly increased mortality), and the effect of temperature on food uptake and growth.

An example of the effect of water temperature anomalies on the growth of the herring biomass is given in Fig. 37-15, showing the sources and sinks of the biomass for a February with average or normal water temperature, and for a February with a 1.5-C positive temperature anomaly. The growth of biomass is considerably enhanced in the latter case, especially in the southern, warmer part of the area. Cold anomalies affect growth less than warm anomalies, as growth is nearly arrested at low temperatures. However, temperature anomalies of cold bottom water have considerable effect on seasonal migrations of flatfishes to feeding and spawning grounds in shallower water. In years with extensive cold bottom water formation on the shelf in the winter, the

spring migrations of flatfishes to shallower water can be considerably delayed (Best, this volume).

Figure 37-11. Distribution of "old pollock" (>45 cm long) in August in the eastern Bering Sea (tons/km²).

Figure 37-12. Change of yellowfin sole biomass distribution due to migration in May and June in the eastern Bering Sea (tons/km²).

Figure 37-13. Change of yellowfin sole biomass distribution due to migration in October and November in the eastern Bering Sea (tons/km²).

Figure 37-14. Distribution of yellowfin sole in August in the eastern Bering Sea (tons/km²).

EFFECTS OF SPACE AND TIME VARIATIONS IN THE DISTRIBUTION OF PREDATOR AND PREY

Predation in any marine ecosystem is largely controlled by predator-prey size relations and the availability and suitability of prey. Thus if some "normal" or preferred prey items are not available, other food items of comparable size are substituted. This process is included in the DYNUMES model. In areas and times when not enough food is available, starvation will occur with several of its consequences, the first of which is usually a delay of sex products development.

Some of the dynamic aspects of predator-prey "overlap," pertaining especially to benthos and to seasonal migrations of flatfish, have already been presented (see Fig. 37-2). The effects of spatial distribution of different prey items on the composition of food of a predator are schematically shown in Fig. 37-16, which depicts a vertical section with predator-prey distribution. Not only does the food composition of the predator vary in space, but the predation pressure on the prey varies as well.

The various possible types of responses of predator and prey biomasses are shown in Fig. 37-17; although these responses can be classified into some defined types, there is considerable overlap of these responses in nature due to various predator-prey dynamic

Figure 37-15. Effect of temperature anomaly on the source and sink of herring in February in the eastern Bering Sea (A: normal February. B: February with a +1.5 C temperature anomaly.).

Figure 37-16. Schematic presentation of the predation by one predator on three prey items with different spatial distributions (presented as a section).

processes (i.e., changes in space and time). Some of the response types are as follows: when the predator biomass decreases (e.g., because of outmigration), local prey increases in abundance (type A), and when predator biomass increases, prey decreases (type B). When the secondary prey is predator to primary prey (i.e., secondary prey is competitor to primary predator), the predator biomass decreases (types C and F).

Figure 37-17. Types of responses to biomass changes in predator-prey controlled ecosystem.

If the prey is mobile and decreases (e.g., by outmigration), the predator biomass might decrease as well (e.g., because of starvation or forced outmigration in search of food) (type D) and vice versa (type E).

The percent of mean monthly zooplankton standing stock consumed per month in the eastern Bering Sea is given in Fig. 37-18. Zooplankton and benthos are the main food resource buffers in the marine ecosystem. The monthly mean zooplankton standing crop was simulated in the model, using quantitative knowledge of its past abundance and seasonal changes. The simulated zooplankton standing stock was made to vary spatially and temporally between 400 and 800 mg/m³. The areas of high zooplankton consumption changed from month to month as affected by the distribution of consumers. This spatial and temporal change of high consumption would allow replenishment by growth and advection in previously heavily grazed areas. It can be assumed that fish eggs and larvae are consumed at the same rate as zooplankton. Thus, if a high zooplankton consumption area coincides with high abundance of pelagic eggs and larvae, low larval survival and low recruitment can be expected of the given species. Since the zooplankton consumption in the northern part of the area is low, a great part of the deceased zooplankton can settle to the bottom, providing detrital food for the abundant benthos. The consumption of benthos is also low in the northern part of the area, although its standing crop is high ("accumulation of generations" in the sense of Spärk). Thus the benthic biomass (or zooplankton biomass) cannot be proportional to the production of fish in all areas.

The consumption of zooplankton near the shelf edge is fairly high (see Fig. 37-18) and this raises several questions about the possible source of zooplankton there. First, the zooplankton simulation in the model for this area might be too low despite the fact that considerably higher quantities were simulated than indicated in the literature, because present quantitative zooplankton catching methods are obviously deficient (Laevastu et al., 1976). Second, the zooplankton, especially euphausiids, might be transported by currents to the convergence at the continental slope.

EFFECTS OF MAMMALS AND BIRDS AS APEX PREDATORS ON THE ECOSYSTEM

During summer the Bering Sea contains more mammals per unit area than any other ocean area. Furthermore, during summer there are more marine birds in the region (over 40 million) than in the rest of the northern hemisphere. Both mammals and

Figure 37-18. Percent of monthly mean zooplankton standing stock consumed in February and in August in the eastern Bering Sea.

birds are apex predators, their effect on the ecosystem being similar to the effect of fishing by man.

In addition to the fact that the occurrence of mammals is seasonal and their distribution uneven, their mobility has various temporal and spatial dynamic effects on the rest of the ecosystem. Estimates of numbers of marine mammals present vary considerably from one source to another. It is apparent that some of the estimates have little to do with reality.

There are three types of seasonal occurrence of mammals—the winter visitors (e.g., “ice” seal and bowhead whale), summer visitors (e.g., fur seal, sea lion, and sperm whale), and year-round residents (e.g., beluga whale).

The estimates of daily food requirements of marine mammals are also variable in the literature. The more conservative estimates are between 4 and 8 percent body weight daily, depending on the species, and these estimates were used in the model.

Seasonal changes of some marine mammals and birds are shown in Fig. 37-19; these estimates are believed to be the best and most realistic estimates, found in various more reliable sources in the literature.

The consumption of fish by marine mammals and birds in the area is about 3 million mt a year, which is about twice the total area catch by the fishery of all nations. The effect on the ecosystem of consumption

by marine mammals is similar to the effect of the fishery—continual removal of biomass without providing much return in the form of food for other biota in the ecosystem. Considering the magnitude of consumption by mammals in relation to harvest by

Figure 37-19. Monthly numbers of some mammals and birds in the eastern Bering Sea.

the regulated fishery, it becomes obvious that fisheries management measures will have a limited effect on fishery resources without simultaneous management of marine mammals as well. Furthermore, the seasonal and selective predation by many mammal species has considerable adverse effects on several fishery resources important for man (e.g., the predation on returning salmon by the beluga whale, fur seal, and sea lion).

VERIFICATION AND VALIDATION OF ECOSYSTEM MODEL RESULTS

Verification of the ecosystem simulation model (DYNUMES III) has been accomplished by ensuring that the formulas used in the model reproduce known effects and behaviors. Further verification has been done by simulating events which are known to produce given changes in the ecosystem. Thus the evaluation of sensitivity in large ecosystem models becomes a continuing study of the response of the ecosystem to changes in various rate and state parameters.

Although the dynamic aspects of the Bering Sea ecosystem are difficult to validate empirically because

(Granfeldt 1979b). Examples of this validation are given in Table 37-2 for the species which are most reliably reported quantitatively in the resource surveys. In general, the results of resource surveys are considered reliable only to about ± 50 percent (Grosslein 1976), whereas the error in the model computation results does not exceed ± 30 percent of the reported value.

Qualitative validation of the simulation models can be provided by occasional special fisheries surveys. In the early stage of the Bering Sea ecosystem modeling, it became obvious that there must be considerable numbers of pollock (and some other fish species) over the deep water in the Bering Sea. However, pollock were never caught over deep water and the model results were severely criticized until a recent Japanese survey showed considerable numbers of older (larger) pollock over deep water. Furthermore, the deep-water areas turned out to be the source of biomass of many pelagic and semipelagic species at least part of the year. The abundant euphausiids in this area provide ample food. However, no extensive schooling occurs over deep water, making the fishery less profitable there than over the continental shelf.

TABLE 37-2

Comparison of exploitable biomasses of some species in the eastern Bering Sea as obtained by surveys and as computed with PROBUB Model (in 1,000 mt)

Species/groups of species	Mean 1975, 1976 surveys (converted) from Bakkala and Smith (1978)	Equilibrium exploitable biomass from PROBUB Model	Catch 1975
Greenland turbot, halibut	176	222	65
Flathead sole, arrowtooth flounder	206	377	26
Yellowfin and rock sole, Alaska plaice	2,716	509	74
Pollock	3,698	6,449	1,285
Cod	233	746	57

of the absence of data, particularly from extensive time-series and behavioral studies, it is possible to validate general abundance and distribution results by comparing the computed abundance and distribution of biomasses with independently obtained empirical data such as those obtained from fisheries surveys. This comparison is rewarding only provided the latter are properly converted using catchability coefficients

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