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BASIC ACOUSTIC OCEANOGRAPHY

APRIL 1975

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ABSTRACT

A brief introduction is given of oceanography for acousticians. The main topics and related acoustical examples are: submarine geology (bottom acoustic loss); marine biology (the deep sound scattering layer); chemical and physical properties of seawater (sound speed and sound absorption); the steady state of the ocean (sound speed vs depth structure and modes of sound propagation in the ocean); and waves in the ocean (sea surface loss and sea surface noise, and acoustic propagation and internal waves).

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FOREWORD

This publication has resulted from a 2 1/2-hour lecture with the same title that the author has presented as part of the Catholic University of America summer graduate course in Ocean Acoustics under the directorship of Dr. Frank A. Andrews. Its purpose is to introduce the vast subject of oceanography in an elementary and compact way, emphasizing those features that would be useful and interesting for acoustic applications. The term "acoustic oceanography" as used here includes studies of the ocean using acoustics as a tool. However, the extremely limited scope of the lecture has led to the inclusion of only one or two acoustic examples under each chapter heading.

J. E. AYRES
Captain, USN
Commander

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1. Introduction

Topics chosen for presentation in a single lecture on the study of the ocean using acoustic signals must be carefully selected and treated briefly. The ocean is mostly opaque to light and radio waves, so that acoustic and other mechanical probes become extremely important in gaining information about the insides of the ocean. The generalized view taken here of acoustic oceanography is one of many possible approaches.

The oceans cover almost 71 percent of the Earth's surface and have a mean depth of 3,800 meters. Some ocean trenches are deeper than the height of the highest mountain on earth. In this report submarine geology is treated first. Natural processes occurring beneath the ocean floor influence the shape and structure of the floor. Sedimentary processes also modify it. The resulting morphology affects acoustic propagation when the ocean floor is involved. Next, marine life is studied in terms of interaction with the ocean environment. Use of acoustic techniques has revealed some interesting habits of sea life, of which one example is the diurnal migration of the deep sound scattering layers. The section on chemical and physical properties of seawater is slanted towards speed of sound and absorption of sound. Although the ocean is a variable medium both temporally and spatially, the next section describes mean state with regard to temperature, salinity, and currents. Acoustic effects caused by stratification of the ocean, as represented by the surface mixed layer, thermocline, and nearly-isothermal cold deep ocean, are treated in terms of the surface duct and deep sound channel. Examples are given of the four chief paths of acoustic wave propagation using ray schematics. The final topic concerns ocean surface waves and internal waves and their statistical variability. Acoustic effects related to the ocean surface, such as noise spectra and sea surface loss, are also treated briefly.

2. Submarine Geology

Submarine geology concerns topography of the sea floor, composition and physical character of sedimentary and igneous materials found on the ocean bottom, and processes involved in the development and modification of topographic relief.

a. The Earth Beneath the Oceans

Earth is a sphere of 6,371-km radius and is considered to be composed of four idealized concentric spherical layers: crust (0-33 km), mantle (33-2,900 km), liquid outer core (2,900-5,100 km), and inner core (5,100-6,371 km) (fig. 2-1).

The boundary between Earth's crust and the denser mantle at an average depth of 33 km is called the Mohorovicic discontinuity, or the Moho for short. The Moho lies at a greater depth under land masses

(20-50 km) than under oceans (5-10 km). Earth's crust consists of several large rigid plates or blocks that float on the more plastic mantle. These plates move relative to one another and most, if not all, earthquake and volcanic activity occurs along plate boundaries. Mid-ocean ridges represent boundaries of plates that are spreading apart. Peripheral mountain ranges, volcanic arcs, and deep-sea trenches represent the boundaries of adjacent plates that are converging (fig. 2-2). When the strength of the Earth's crust is occasionally exceeded locally, sudden ruptures occur that displace rock masses along fractures called faults. Most faulting occurs at the borders of crustal plates.

Many Earth scientists believe the lateral spreading apart of crustal plates is caused by motion of convective cells in the mantle (fig. 2-3). Molten material in these cells rotates about cell axes, forcing hot basaltic mantle material up through and along the underside of the crust, slowly dragging the crustal plate until a zone of downward cell motion is reached. The Mid-Atlantic Ridge is estimated to be spreading on the average of 1.25 cm/yr, whereas the East Pacific Rise is spreading at a rate four times as great. Since new crust is forming, old crust must be destroyed somewhere because the earth is not growing larger at the required rate. A process of subduction occurs at trenches, which are depressions where crustal plates are forced, or sink, into the underlying mantle. Slippage of material into trenches gives rise to earthquakes and volcanic activity, which leads to formation of seamounts and island arcs.

b. Bathymetry

Most detailed measurements of ocean depths and submarine topography have been made in the last 50 years, since the development of echo-sounding equipment. This acoustic equipment measures travel time of a sound pulse to the sea bottom and return, producing a continuous trace of the profile of the sea floor along a ship's track. Figure 2-4 shows a bottom profile of a vertical section of the North Atlantic Ocean along a track between Cape Henry, Virginia, and Rio de Oro, Africa. The continental margin of each continent consists of a shelf, a slope, and a rise. Ocean basins are separated by the Mid-Atlantic Ridge. The old volcanic island of Bermuda is also seen.

After collecting and assembling appropriate bathymetric data from many such tracks, detailed bottom contour charts are prepared which may be used for navigation as well as for other scientific purposes. Figure 2-5 is a bottom topography chart of the North Atlantic Ocean. This chart shows the lateral extent of the major physiographic provinces of the ocean floor.

The hypsographic curve (fig. 2-6) is a plot of the cumulative areal distribution of heights above and depths below mean sea level. This curve is used to show the area of different sea floor zones of the Earth's oceans. A shallow zone at depths from 0 to 200 m occupies

about 7.6 percent of the total ocean area. In this zone are the continental shelves, the shallow aprons surrounding continental land masses. Continental shelves vary greatly in width, from a few miles off the California coast to some 800 miles in the Siberian Arctic. Although the average slope of the shelf is gentle, within the shelf boundaries, terraces, ridges, hills, depressions, and steep-walled canyons may occur.

Next, a deeper zone of the continental margin from about 200 to 3,000 m deep takes up about 15.3 percent of the oceanic area. Within this zone are the continental slopes of the oceanic floor, regions in which continental land blocks drop off rapidly to the deeper sea floor. A common feature of the slope is the occurrence of extraordinarily large gullies, known as submarine canyons, which are steep-walled fissures penetrating the slope and cutting across the shelf.

The continental slopes, with a typical gradient of about 4° , drop smoothly to the incline of the continental rise which has a gradient of about 0.3° to 0.05° . The continental rise then merges into the floor of the ocean basins.

The greatest portion, 75.9 percent of the sea floor, lies at depths ranging from 3,000 to 6,000 m, the zone of large oceanic depressions called basins. These are usually considered to be the Atlantic, Pacific, and Indian. Other large areas also are sometimes considered as separate units, such as the North Polar Basin, North American Basin, and Pacific-Antarctic Basin.

A major feature of the ocean floor is the mid-ocean ridge about 1,500 to 2,000 km wide that extends for 60,000 km and covers about 25 percent of the ocean floor, an area about equal to that of all the continents. Throughout most of its length, the ridge is split by a deep rift in which many earthquakes originate, the ridge being the locus of a crack in the Earth's crust. The mid-ocean ridge runs nearly twice around the Earth, occupying about the same position on the hypsographic curve as continental slopes.

Only 1.2 percent of sea floor lies at depths ranging from 6,000 to 11,000 m. This is the area of troughs, trenches, and deep basins. The deepest trenches are found in the Pacific Ocean and exceed 10,000 m, whereas the deepest point in the Atlantic Ocean is the Puerto Rico Trench at 8,750 m. Note that the depth of the deepest trench is greater than the height of the highest mountain.

c. Bottom Sediments

The crust of the ocean floor is coated by a sedimentary blanket that has buried or modified the original igneous surface. The morphology of the sea floor is, therefore, the result of generation of igneous topography through processes of faulting and volcanism at the crest of the mid-oceanic ridge, lateral motion away from the ridge crest in a conveyor belt fashion, and slow burial by sedimentary processes.

Regional elevation depends on age of the particular parcel of crust. The original igneous crust is mantled by sediments into zones reflecting age of the basin, proximity to continents and sediment sources, and areas of biologic productivity. The three primary sedimentary processes for smoothing of the original igneous rock are: (1) continuous pelagic "snowfall"; (2) gravitational slumps and slides, including turbidity flows from both terrigenous sources and, on a smaller scale, the elevated undersea peaks; and (3) redistribution of sediment by bottom currents.

Figure 2-7 shows the geographic distribution of deep sea sediments by source as explained below. Sediments may be classified in increasing size order as clay, silt, sand, and gravel. Sediments are also classified by source.

(1) Terrigenous -- Organic or inorganic land-derived materials; material derived from inorganic rock is lithogenic.

(2) Biogenic -- Organic or inorganic material from marine organisms.

(3) Volcanic -- Submarine or airborne volcanic material (lava, ash).

(4) Hydrogenic -- Materials crystallized from seawater, such as manganese nodules.

(5) Cosmic -- Materials of extraterrestrial origin such as cosmic dust and tektites.

Sediments are often classified by the biotic form that produces their major biogenic component. These sediments are called oozes. Microscopic free-floating plants and animals of the sea leave hard remains or tests of either a calcareous or siliceous nature. Calcareous deposits, though widespread, are restricted to shallower regions of the pelagic zone than siliceous deposits, since calcareous remains are more soluble than siliceous remains. Brown or red clays predominate in deeper parts of the ocean basins under surface waters of low biological production. These clays are derived from the very smallest particles of wind-carried atmospheric dust from land.

Coring of the sea bottom and collection of material by use of dredges and grab samplers are techniques for obtaining samples from the upper layers and surface of the sediment. These methods are slow, laborious, and expensive. Deep sea photography is also used for studying the characteristics of the sea bottom. However, seismic profiling in the deep ocean, introduced for routine shipboard use in 1956, has become one of the most widely used tools of marine geologists and geophysicists. Seismic profiling is a method of studying sediment thickness and layering, using the penetrability of low-frequency acoustic energy into the bottom (fig. 2-8). If the acoustic energy density is

high enough, the energy remaining after partial reflection at the sea floor penetrates sediments and suffers partial reflections at each successive boundary. The bedrock under the sediments also reflects energy, marking the bottom of sedimentary material. Sound travels at a determinable speed in water and in sediments; the time interval between the outgoing pulse and return of the reflected energy then determines distance between reflecting horizons. Subbottom profiling yields information on the sediments, topography, and structure of the oceanic crust. For determining crustal thickness with depths of many kilometers, sound in the frequency range of 3 to 30 Hz is used; whereas for measurements of sediment thickness of the order of tenths to a few kilometers of sediments, seismic reflection profilers in the 30- to 500-Hz range are used. For studying the detailed stratification of the upper sediments, in the top 100 m, echo-sounders in the range 3.5 kHz to 12 kHz are used. As can be inferred from the previous information, ocean bottom acoustic characteristics are strongly frequency or wavelength dependent.

From seismic studies it has been found that the average thickness of the crust including the ocean sedimentary layer, down to the mantle is 6.0 km, compared with an average crustal thickness of 33 km on land. The average thickness of the present oceanic consolidated or compacted sediments is 3 km. This is equivalent to 9 km of unconsolidated sediment, which still has interstitial water in its pore spaces. The thickness of sediments and rates of sedimentation are used to estimate the age of an oceanic basin, with 200 to 600 years considered the average time for settling to a depth of 2,900 m. This corresponds to a particle size of about 4×10^{-5} cm, which is in the silt-size range. Of course, this is a grossly oversimplified model.

d. Bottom Acoustic Loss

The loss of acoustic energy impinging on the bottom is quite important for long-range sound propagation in the ocean, as well as for understanding subbottom structure. Acoustic energy may be lost by partial transmission into the bottom and subsequent absorption or by scattering into nonspecular directions due to bottom roughness. Bottom reflection loss can vary considerably from one location to another.

Important parameters are the sound speed and density contrasts between water and sediment layers in the bottom as well as porosity of the sediments. Thicknesses of the bottom layers must be considered also, especially in terms of acoustic wavelength or frequency. For sediments with porosities greater than about 65 percent, sound speed in the bottom is slower than in the deep bottom water. For such sediments there is an angle of intromission, at about 10° grazing angle, with very large losses of about 26 dB. The losses then decrease to about 14 dB from 40° to 90° grazing angles. On the other hand "fast" bottoms, with sound speed greater than water, will show a critical grazing angle at about 30° , with practically 0 dB loss from 0° - 30° grazing angles, and a sharp rise with losses of about 15 dB from 60° to 90° (fig. 2-9).

3. Marine Biology

a. The Sea as a Biological Environment

The sea has several distinct advantages over the land for the development of life. Water, necessary to all life, is immediately available. Beyond this, seawater contains in solution the required gases and other materials for growth, and is very similar in composition and solute concentration to the body fluids of all animals. The resulting near equality of the osmotic pressure of seawater and body fluids permits the direct intake of nutrients and excretion of waste in some organisms. Buoyancy provided by the high density of seawater reduces the need for skeletal structure, whereas the high specific heat of seawater reduces the possibility of rapid temperature change. The range of pressures experienced in the ocean as a whole is very great, and many organisms are consequently restricted to specific layers, but all life is not excluded from the deep abyssal regions. Bathyscaphe dives, in particular, have confirmed the existence of marine life in the deepest parts of the ocean.

Overall, a considerable variety of conditions exists in terms of salinity, temperature, illumination, and pressure in the ocean. Great variability with time is largely confined to coastal waters and shallower layers. Here, marine forms must be adaptable to changes or capable of adjusting themselves to the optimum condition. Light in the surface layer is particularly important for photosynthesis by floating microscopic plants. Notably, there is an adjustment to lighting conditions by some marine life, resulting in a diurnal migration toward and from the sea surface. Large populations of this marine life can form sound scattering layers. As distinct from regions where marked variability exists, large volumes of deep water have remarkably constant conditions, and specific forms appropriate to the particular environment can develop. Circulation processes in the sea are important to life in many ways. These processes maintain oxygen content, replenish the supply of nutrients, and aid in dispersal of waste material and young life in the form of spores, eggs, larvae and young adults. Clearly, however, unexpected and unseasonable disturbances of the circulation pattern can, by moving life into an unacceptable environment, result in considerable destruction.

Formally, the marine biological environment is divided into the water region, called pelagic, and sea floor region, referred to as benthic. The water region is subdivided horizontally into the neritic zone, which lies over continental shelves (low tide to 200-m depth) and the oceanic zone, which lies seaward of the shelves. The oceanic zone is further divided into depth regions: epipelagic, from the sea surface to about 200-m depth; mesopelagic, 200 m to 1,000 m; bathypelagic, 1,000 to 4,000 m; and abyssopelagic (hadal), the deepest parts of the ocean (fig. 3-1).

The most significant biologic zonation is the division into the euphotic or photic zone, in which there is sufficient solar radiation for active photosynthesis, and the deeper aphotic region. The photic zone may extend to about 180 m in clear oceanic water at low latitudes, or may be as shallow as a few meters in the more turbid coastal regions of temperate latitudes (fig. 3-2).

There is considerable seasonal change in surface solar radiation in the temperate zone; therefore, thickness of the photic zone varies throughout the year. In the spring, when radiation levels increase, plant life grows rapidly, using nutrients supplied to surface waters from depth by winter mixing. Surface heating in spring imposes a stability on the water column that reduces vertical mixing and increases the residence time of phytoplankton in the layer above the developing thermocline. Retention of phytoplankton near the surface, increased radiation, and available nutrients trigger the spring bloom. In the tropics, low nutrient supply prevents high productivity of organisms; in temperate regions, both light and nutrients may limit production; and in polar regions, light alone controls productivity. Nutrients are supplied to the photic zone by: (1) rivers which supply land material, and (2) upwelling of water from depths below the photic zone where nutrients have accumulated due to decomposition. In the sea, as on land, photosynthetic plant life must exist in order for animal life to survive.

b. Organisms and the Composition of Seawater

Seawater is more than a solution of salts in a certain proportion, for marine life is an inherent part of the sea, and marine organisms play an important role in determining the chemical and physical properties at any particular location and time, e.g., they have an effect on oxygen and carbon dioxide gaseous contents. Beyond this, certain elements occurring in small concentrations in seawater are removed in relatively large amounts by marine life for skeletal, structural, and other uses. Inorganic compounds of nitrogen, phosphorus, iron, and silicon -- the plant nutrients -- fall in this class, and their distribution is affected appreciably by biological activity.

Life cycles of the various organisms in the sea are closely interwoven. Plants of all sizes furnish the basic food for all life and grow in the upper illuminated layer. These plants absorb carbon dioxide from solution and release oxygen. They also absorb phosphates and nitrogen salts and convert them into organic matter. However, the inorganic substances that they use for growth must be made available to them near the surface, in spite of downward movement of particulate matter in the sea under the influence of gravity. Availability of these materials at the surface results from circulation of ocean waters, convective mixing, and upwelling from the bottom. The inflow of minerals from land is a small factor in supplying nutrients for sea plants. Whenever plant nutrients are abundant near the surface, large

marine populations occur, as compared with other parts of the ocean. In deep water, such areas can be identified by their color, for water rich in life is greenish, whereas "desert" water is blue.

c. Marine Life

The sea's inhabitants range from a multitude of microscopic one-celled organisms to the 100-ton blue whale. Smaller forms, especially plants, are extremely prolific under proper conditions and furnish the food supply for successively larger forms of animals. In describing the population of the sea, three major groups are distinguished depending upon habitat: nekton, benthos, and plankton. Nekton is the name given to swimming animals that can move freely over large distances. The term benthos refers to the sessile animals such as sponges, creeping forms such as crabs, and burrowing forms such as worms found at and in the sea bottom. The plankton includes all floating plants (phytoplankton), animals that live permanently in a floating state (zooplankton), and also larvae and eggs of animal benthos and nekton (meroplankton). The plant plankton (e.g., diatoms, dinoflagellates, and coccolithophores) and animal plankton (e.g., globigerina and radiolaria) are so numerous that extensive areas of the ocean floor are covered with oozes (fig. 2-7) containing their skeletons. There is some question regarding the grouping of motile dinoflagellates as plants even though they carry on photosynthesis. Figure 3-3 shows a zooplankton sample, including meroplankton.

In the sea, as on land, plants are capable of developing complex organic substances from the simple compounds dissolved in water. Without these plants animal life would be restricted to a small region near the shore where organic material of terrestrial origin exists. Nearly all marine plants are thallophytes, primitive plants with no true root, stem, or leaf. They include marine algae and fungi, and bacteria are frequently included with them. Brown algae vary greatly in size, and plants of this class form the "kelp beds" commonly harvested for commercial purposes. Yellow-green algae are mainly floating forms and include diatoms, which are unicellular in structure, and dinoflagellates, again typically unicellular, but possessing two flagella for locomotion. The latter are second only to diatoms as producers in the marine plankton. Some of the latter have the capacity to luminesce when disturbed. Mosses (bryophytes) and ferns (pteridophytes), two major divisions of the plant kingdom, are not represented in the sea. The highest plants, the spermatophytes, are represented, however, by some 30 species of flowering plants. They have invaded the sea by way of fresh water.

Of more than 50 animal classes, only four are exclusively nonmarine. In contrast to plant life, all major groups (phyla) of animal life are represented in the sea. Each phylum is subdivided successively into class, order, family, genus, and species. When thinking of animal life in the sea, one tends to emphasize the higher

order of marine animals such as the whale and larger fishes, but these are an extremely small part of the total animal life.

d. The Deep Sound Scattering Layer

Volumes have been written on the complex sound production and reception by porpoises and other cetaceans for food hunting, navigation, and communication. Noise making by croakers and snapping shrimp is also covered elsewhere. This section, however, will be concluded with some remarks regarding the acoustic effects of the "deep scattering layer." The deep scattering layer is named from its property of scattering sonar signals and showing up as layers on depth sounding records (fig. 3-4).

Actually, there appear to be two principal deep scattering layers. The shallower layer has an average midday depth of 240 m, with extremes between 185 and 405 m. Average thickness is about 75 m. The deeper of the two layers has an average midday depth of about 500 m, extreme values being 405 and 590 m. Average thickness of this layer is 130 m. The average migration rate at sunrise and sunset of elements in the 500-m layer is about 7.5 m/min and that in the 240-m layer is about 4.5 m/min. The 500-m layer usually is not found in water depths of less than 1800 m.

The food chain of the sea is intricately involved in biological communities forming the scattering layers. A range of plant and animal sizes occurs from microscopic to the largest fishes, all feeding on each other. Scattering layers are sensitive to light, approaching the surface at sunset and going deep at sunrise (fig. 3-4). Levels to which the animals descend appear to be related to their negative response to light and sensitivity to sudden temperature decreases.

Acoustic scattering properties of deep scattering layers are similar to those expected from a collection of simple air bubbles in water and show resonant peaks as a function of size and depth (fig. 3-5). In fact, bubbles scatter sound much more strongly than solid particles of the same size.

The main scatterers are fish with gas-filled swimbladders. To maintain their constant size, these fishes are able to generate gas in their swimbladders as they descend and vent gas through their gills via the blood stream as they ascend. As verified by photographs and net trawls, most numerous mesopelagic fauna, with swimbladders, between depths of 150 and 1,000 m are myctophids or lantern fish, bristle-mouths, and hatchetfish (fig. 3-6). Myctophids reach a maximum length of from 4 to 12 cm, with swimbladder diameters of about 4 percent of their length. Resonant frequencies are observed between 3 to 20 kHz, corresponding to fish lengths of from 10 cm to 1 cm. It is interesting that siphonophores have a small carbon monoxide bubble in their float and scatter 12-58 kHz at 100 m depth and 24-111 kHz at 400 m depth.

Some indication of regions of relative sound-scattering strength in the ocean may be obtained from figure 3-2, which shows the geographic distribution of transparency, since transparency in the open ocean would be affected by the density of plankton as well as solid particles. The Sargasso Sea region, for example, is exceptionally clear, whereas the upwelling regions near Iceland are rated only clear.

4. Chemical and Physical Properties of Seawater

a. Composition

Seawater is a dilute solution of salts. Its composition in terms of ion concentrations is given in figure 4-1. Four ions (Na^+ , Mg^{++} , Cl^- , and SO_4^{--}) constitute 97 percent of the total. The formal definition of salinity follows: "Salinity is the total amount in grams of solid material dissolved in 1 kg of seawater when all carbonate has been converted to oxide, all the iodine and bromine have been replaced by chlorine, and all organic matter has been completely oxidized" (fig. 4-2). It is expressed in grams of salt per kilogram of seawater (parts per thousand by weight of seawater is represented by the symbol ‰). An empirical relationship between salinity and chloride content (chlorinity) is used to convert a measurement of chloride content to salinity,

$$S \text{ ‰} = 1.80655 \text{ Cl ‰}. \quad (S = 35.00 \text{ ‰ for Cl} = 19.37 \text{ ‰})$$

Proportions of the different constituents are nearly constant in the open sea, although the total salinity varies between about 33 and 38 ‰, depending upon precipitation, evaporation, and ice melting. Considerably lower salinities may be observed in coastal regions and in almost-closed seas with low evaporation and a considerable inflow of fresh water, e.g., the Baltic, where the salinity may be as low as 5 to 8 ‰. Conversely, with little inflow and high evaporation, as in the Red Sea and Persian Gulf, salinities reach as high as 40 ‰. Nevertheless, the assumption of a mean salinity of 35 ‰ for seawater (90 percent of the ocean differs by less than 3 percent from this value) is adequate for many applications where high precision is not required and to define many of its physical properties.

Trace quantities of at least 50 of the elements have been identified in seawater. Surprisingly, the sea is very poor in iron. Significantly, copper replaces iron in the blood pigment of many marine animals. Sea animals and plants sometimes concentrate trace elements in their systems way out of proportion to the natural concentration. For example, seaweeds can concentrate in their tissues the small amount of iodine in the sea to the point where they may be harvested to obtain commercial quantities. In some fish, concentration of heavy metal pollutants in their systems makes them poisonous to eat. Considerable attention has been given in recent years to the radioactivity of seawater, the radium content averaging about 0.07×10^{-12} ‰, which is about 10^{-4} of that in rocks of the Earth's crust.

Gases may enter solution either by purely physical absorption or chemical combination with the liquid. For the ocean-atmosphere system, physical absorption occurs in the case of oxygen, nitrogen, and rare gases. Carbon dioxide, on the other hand, reacts with water and dissolved salts. In comparison with the oxygen-nitrogen ratio in the atmosphere of 1:4, that of air dissolved in water is more nearly 1:2. Saturation quantities are a function of water temperature and decrease by 50 percent from polar to equatorial regions. There is ample air available at the sea surface, therefore, gas contents in the upper layers are near saturation. As nitrogen is relatively inert, only minor variations with depth are observed, since all water has been in contact with the atmosphere at some time. The oxygen content is markedly affected by biological factors, such as generation by photosynthesis in the upper layers, which occasionally results in supersaturation.

Oxygen consumption by animal respiration and bacterial oxidation reduces the oxygen content and results in a pronounced minimum at depths of 100 to 1,500 m in the middle and lower latitudes. Deeper waters of polar origin are relatively rich in oxygen. In stagnant water, where there is no photosynthesis and oxygen depletion occurs, as in the deep portion of the Black Sea, hydrogen sulphide may be formed. The partial pressure of CO_2 in the sea in contact with the atmosphere tends toward equilibrium with that in the air. However, areas exist where carbon dioxide-rich water is brought to the surface by rising currents, while in other regions the CO_2 partial pressure is frequently low in spring and summer where plant plankton is assimilating. Figure 4-3 is based on the condition that the volume of the atmosphere is about 2.9 times that of the ocean. Figure 4-3, comparing gas concentrations in air and seawater, shows that the ocean is a sink for CO_2 .

The pH is a measure of hydrogen ion concentration on a logarithmic scale. The pH of pure water is 7. An acidic solution has a pH of less than 7, whereas an alkaline solution has a pH greater than 7. Seawater is normally alkaline with a pH between 7.5 and 8.4.

b. Equation of State and Thermal Properties

The equation of state represents the density of seawater as a function of salinity, temperature, and pressure. This equation is of particular importance in oceanography in that minor changes in density affect stability, water mass transfer, and turbulent mixing. For these purposes, the oceanographer must know the density to 1 part in 10^5 and must resort to indirect measurements of chlorinity, electrical conductivity, refractive index, or the like for density determination, with corrections for depth and temperature. The expression for density is unwieldy and in practice it is customary to use the notation $\sigma_t = (\rho - 1) \times 10^3$ where ρ is in g/cm^3 , and to use tables of σ_t as a function of salinity, temperature, and pressure.

Density of seawater at the surface of the open sea ranges from about 1.021 to 1.027 g/cm³. Generally about 1.025 g/cm³, density increases with increasing salinity and continually decreases with increasing temperature, provided the salinity is greater than 24.7 ‰. For lower salinities there is a temperature of maximum density as for pure water. A small increase of density occurs with increase in pressure, which is important because of high pressures in the deep ocean.

Seawater exhibits the unique properties of solutions, namely, osmotic pressure, depression of freezing point, elevation of boiling point, and reduction of vapor pressure. The observed freezing point is -1.910°C for a salinity of 35 ‰, the freezing point depression being less than that computed on a basis of complete ionic dissociation by a factor of 0.89. For all practical purposes, the latent heat of evaporation may be taken as that of pure water.

Specific heat at constant pressure C_p is close to 1 cal/g°C and increases with increasing pressure. The ratio of specific heat at constant pressure to that at constant volume increases with temperature and pressure. Seawater is compressible, and effects of adiabatic processes, although small, must be taken into account when studying the vertical distribution of temperature in the great depths of the oceans. Adiabatic processes are those that take place without the flow of heat into or out of the volume of the medium under consideration. When a sample of water is brought in an insulated sampling bottle from an initial depth (pressure) to the surface, a certain amount of thermodynamic cooling occurs because of the change in pressure, irrespective of the temperature of the local environment. The temperature that the sample would have when brought from the initial depth to the surface without heat flow is called the "potential temperature."

In the ocean, the important mechanical mixing property is eddy viscosity, which is the effective viscosity occurring on a macroscopic scale in large bodies of fluid. This concept is to be differentiated from molecular viscosity, which is measured on a small sample of fluid in the laboratory. Dynamic viscosity is defined as the ratio of shearing stress to the resulting velocity gradient and expressed in grams per centimeter per second. It is common in fluid mechanics to use kinematic viscosity, which is dynamic viscosity divided by density and expressed in square centimeters per second. The molecular viscosity for water is about 0.019 cm²/s at 0°C and approximately half this at 20°C; the eddy viscosity is substantially larger and may range from 2 to 7,500 cm²/s depending upon conditions. Surface tension controls the capillary waves from which larger surface waves receive energy from the wind. The surface tension value is about 76 dyn/cm at 0°C and 73 dyn/cm at 20°C.

c. Optical and Radio Properties

Reduction of the intensity of optical radiation in the sea is of major biological interest. In general, light is confined to the

upper few hundred feet of the ocean. Extinction of light is due to both absorption and scattering, the former being greater at the red end of the spectrum and the latter at the blue. Extinction is largely a function of the amount of suspended material and is considerably greater in coastal waters. The refractive index of seawater depends strongly on salinity and is used to measure salinity (with appropriate temperature corrections). Electrical conductivity rises rapidly with an increase of salinity, and as a result electrical conductivity measurements are used also for the determination of salinity. (Again, there is a strong temperature dependence and a correction must be made.) Conductivity, which is approximately 4 mho/m, limits the transmission of radio waves. Figure 4-4 compares absorption of radio waves and sound waves in seawater. It may be seen that sound is transmitted better below about 100 MHz. As seawater has a relative permittivity of 80 for radio waves, it is a conductor below 10 MHz and a dielectric above 10^5 MHz. The skin depth δ at which radio waves decay to $1/e$ of their surface field intensity value is given by $\delta = 250/\sqrt{f}$ meters, where f , the frequency, is in Hz. At 1 MHz, $\delta = 0.25$ m. Ten kW of power in a plane wave at 16 kHz is attenuated to $1 \mu\text{W}$ at a depth of 23 m. The speed of radio waves is reduced considerably by the conductivity, being equal to $2\pi f \delta$. Thus, at 1 MHz, speed equals 1.5×10^6 m/s and is 5.0×10^4 m/s at 1 kHz.

d. Sound Speed

Of the many acoustic parameters associated with the ocean and its boundaries, the most important are speed of sound and absorption loss.

The speed of sound, C , in seawater is given by the formula $C^2 = (\partial p / \partial \rho)_S$, where S denotes an adiabatic process. Sound speed is related to the equation of state where ρ , the density, is given in terms of pressure, temperature, and salinity. An equivalent form for the speed of sound is $C^2 = 1/(\rho K_a)$ where K_a is the adiabatic compressibility. The adiabatic compressibility is the reciprocal of adiabatic bulk modulus. There are several expressions for the speed of sound in terms of temperature, salinity, and pressure. The formula most widely accepted is that of Wilson shown in figure 4-5.

At 0°C , the speed of sound is 1449.3 m/s for a salinity of 35 ‰ at atmospheric pressure and increases by about 1.4 m/s per 1 ‰ increase in salinity, by 4.5 m/s per $^\circ\text{C}$ increase in temperature, and by 0.016 m/s increase in depth of 1 m (1 atmosphere $\approx$ 10 m). These variations with depth can result in marked acoustic refraction in the ocean.

e. Sound Absorption

Absorption of sound in seawater increases rapidly with acoustic frequency and is dependent on salinity, temperature, and pressure.

Absorption values are approximately 0.065 dB/km at 1 kHz and 1 dB/km at 10 kHz ($T = 4.0^{\circ}\text{C}$). The absorption coefficient is that component of the attenuation coefficient that passes directly into heat in the ocean.

Absorption of sound in pure water, computed from Rayleigh's formula based on shear viscosity, is about one-third of the measured absorption. The water molecule has a bent-triangle structure, with an angle of 109° between the oxygen and two hydrogen atoms. Two modes of packing exist for water molecules, ice and liquid, both of which are always present in varying relative amounts in liquid water. The density maximum at 4°C and atmospheric pressure for the liquid is due to a relative maximum amount of the closely packed structure. Molecular packing is pressure dependent, and in the presence of a compressional sound wave, there is a relaxation effect causing the packing of some molecular aggregates to shift back and forth between liquid and ice structure. The extraction of acoustic energy in this process is the pure water absorption.

In seawater, the additional absorption over that of pure water is due to a pressure sensitive relaxation process at about 70 kHz, arising from a shift between hydrated and nonhydrated magnesium sulphate ions (fig. 4-6). Magnesium sulphate accounts for 5 percent of the salts present in seawater. The effect of sodium chloride, accounting for 78 percent of seawater salinity, is to inhibit absorption. An expression for the absorption of sound in seawater for frequencies above 2.5 kHz as a function of temperature, salinity, and pressure is given in figure 4-7.

A relaxation occurring at about 1 kHz increases seawater absorption further, by a factor of 10 over the magnesium sulphate absorption, at very low frequencies below 1 kHz. This relaxation process is due to the boric acid-borate system in seawater. Thorp's empirical expression for absorption of sound in seawater over the entire frequency range is given in figure 4-8.

5. The Steady State of the Ocean

a. Heat Transport by Winds and Currents

Since the average temperature of Earth as a whole is not changing appreciably, then the total amount of heat received from the Sun must equal the amount lost by radiation back into space. However, at any one time and place, Earth is either warming or cooling. The intensity of solar radiation is largely a function of latitude and season.

Figure 5-1 shows the radiation balance between insolation and outgoing radiation as a function of latitude. The surplus in the tropical regions is balanced by the deficit in polar regions. The

resulting latitudinal temperature distribution is shown to vary from about 28°C at the equator to about -1.7°C at the poles. Much greater differences in temperature between the tropics and poles would be experienced, if it were not for the relatively free circulation of atmosphere and oceans, which compose the working substance of a gigantic heat exchanger. Large-scale heat transport from warmer to colder regions is due to atmospheric winds and ocean currents.

Air, which acts as a major mechanism for redistributing heat, is driven by unequal heating of the Earth's surface to rise at the equator and sink at the poles. However, air cannot circulate uniformly in a single convection cell, since Earth's rotation, through the deflecting force of the Coriolis effect, prevents a direct north-south path for the air. Earth's rotation and the distribution of land and water cause a three-celled vertical circulation system from the equator to a pole (fig. 5-2).

In the oceanic environment, moving air produces waves and drives the major surface currents. Once surface waters are set in motion, they create distributions of sea-surface temperatures which in turn alter the winds that drive the currents. There is, however, a mean steady-state condition.

A wind-driven current does not flow in exactly the same direction as the wind, but is deflected by Earth's rotation (fig. 5-3). This deflecting force, the Coriolis effect, is greater at higher latitudes and is to the right of wind direction in the Northern Hemisphere and to the left in the Southern Hemisphere. Between 10°N and 10°S , the current usually sets downwind. In general, angular direction between wind and the surface current varies from about 10° in shallow coastal areas to as much as 45° in some areas of the open ocean. The angle increases with the depth of current, and at certain depths the current may flow in the opposite direction to that at the surface.

Ocean current patterns, consisting of huge gyres, are related to wind systems of the world (fig. 5-4). Water motion at the surface, modified by Coriolis effect, responds to: (1) density gradients caused by latitudinal heating effects, and (2) surface winds, which themselves are modified by the Earth's rotation and the distribution of land and water masses. Position of the gyres shifts only slightly with season, except in the northern part of the Indian Ocean and along the China coast, where monsoons cause the currents to flow in opposite directions in winter and summer.

Ocean currents may attain speeds of 5 or more knots as reached by the Florida current. In middle latitudes, the strongest surface currents rarely reach speeds above 2 knots. At depth, currents are generally slow, usually less than 0.5 knot. However, there are submerged high-velocity streams, such as the Cromwell Current in the Pacific Ocean.

A slow, density-driven, vertical circulation occurs in the ocean that leads to a stable density layering or stratification. Water is heated in the tropics. This heated water decreases in density and expands, spreads northward and southward along the surface, and gradually cools as it approaches higher latitudes. As warm surface water is drained off from the tropics it is replaced by upwelling cold water that originated in polar regions. The deepwater layer of all oceans is derived from polar regions and has about the same characteristics of salinity and temperature as the surface waters of those regions. This happens because the water is uniformly cold in these regions and, along with brine residue left behind in ice formation, produces the heaviest waters in the ocean. This water sinks and spreads from one pole to the other. The horizontal net drift in these deep thick layers is very small, about 1 to 3 cm/s. This produces a low average temperature about 3.8°C for the whole ocean, and a conservative temperature-salinity relationship.

b. Three-Layered Vertical Temperature Structure

Local convection is important in heat exchange near the surface. Surface waters may be warmed down to 10 meters by solar radiation, and may be cooled by evaporation, precipitation, and radiation. Both cooled water and more saline water will be heavier than that beneath it and will sink. These processes together with wind result in vertical mixing and widespread formation of an isothermal surface layer.

The basic vertical thermal structure is essentially a three-layered system (fig. 5-5): (1) a relatively warm thin surface layer mixed by wind and convection processes, (2) a very deep mass of much colder water in which the temperature decreases very slightly and uniformly with depth, and (3) a transition layer, known as the main thermocline. This structure is an inherently stable one since the density, which is largely controlled by the temperature, increases with depth (pycnocline). The warm surface layer is thickest in mid-latitudes as is the thermocline or pycnocline. In latitudes higher than about 50°N and S, no surface layer exists, and the entire water column is the same temperature as the cold deep ocean water. Superimposed upon this stable situation are marked diurnal and seasonal temperature variations in the upper layers. Illustrating the three-layered ocean, figure 5-6 shows actual plots of temperature-depth curves for three different latitudes for summer and winter.

Figure 5-7 indicates seasonal effects in the near-surface layer due to wind and convective mixing. If the spring condition is a moderately deep mixed layer, then surface heating and calm conditions of summer will create a surface thermocline leaving an isothermal section in the profile (fig. 5-7A). Winds then mix the near-surface water creating a pair of mixed layers and thermoclines (fig. 5-7B). When strong winds blow, one deep mixed layer develops (fig. 5-7C). In autumn, surface cooling (fig. 5-7D) will induce convective mixing (fig. 5-7E), and by winter the mixed layer is very deep (fig. 5-7F).

Figure 5-8 shows frequency of occurrence by percent of mixed layer depths in the North Atlantic by season. In summer and fall they are mostly shallower than 250 feet (76 m). In winter they are mostly deeper than 250 feet.

c. Salinity and Water Masses

Characteristics of ocean waters are controlled by surface processes, for it is only at the sea surface that heat or freshwater is added or removed in quantity. Thus, salinity and temperature of midocean surface waters are determined by the distribution of evaporation and precipitation processes as well as solar radiation. Surface salinities reach maximum in belts centering at about latitudes 23°N and S . Evaporation is maximum along the outer edges of the trade wind belt, where the air is descending and thus relatively dry. More precipitation than evaporation occurs in the tropics and at latitudes above 40°N and S . Salinity in the surface waters of the ocean is proportional to the excess of evaporation over precipitation. This is demonstrated in figure 5-9.

Salinity effects on density are sizable, but the pycnocline is controlled by development of the thermocline because of the relatively large temperature range (-1.7°C to 30°C) of seawater compared with the relatively small salinity range (33 to $37\text{ }^{\circ}/\text{oo}$).

In surface waters, the relationship between temperature and salinity is highly variable because of surface heating and cooling, evaporation, and precipitation. In the main thermocline and deepwater layer, however, a direct correlation between temperature and salinity exists. Knowing the temperature and water mass, one can predict the salinity with almost as much accuracy as it can be measured by the routine process ($\pm 0.02\text{ }^{\circ}/\text{oo}$).

The permanency of the oceans' gross vertical structure, maintained by vertical circulation, allows identification of water masses from their vertical salinity and temperature distributions. If the vertical distributions of temperature and salinity are plotted on a graph with temperature and salinity as coordinates, and density excess (σ_t) as a parameter, then the resulting characteristic curve is called a T-S curve. Figure 5-10 shows T-S curves for the Atlantic Ocean. Figure 5-11 shows the water masses of the world oceans classified from their T-S curves.

d. Sound Speed vs. Depth Structure

The density stratification of the ocean, with the three-layered temperature model, leads to a typical sound speed vs. depth profile such as seen in figure 5-12.

In the thermocline, temperature changes rapidly and dominates the sound speed profile. In the mixed layers, the temperature profile

is isothermal, and in the deep layer it changes slowly. In deep-layer regions, the pressure effect on sound speed prevails. Direct salinity effects on sound speed are usually small.

Figure 5-12 shows that when an isothermal mixed layer exists as in winter, sound speed increases with depth from the surface at the rate 0.016 m/s/m until the temperature effect of the thermocline takes over. Beneath the thermocline, decrease of temperature with depth becomes small, the pressure effect again takes over and sound speed increases again close to the isothermal rate of 0.016 m/s/m. In figure 5-12, another velocity minimum occurs followed by an increase to the highest sound speeds in the ocean near the bottom. Total variation in sound speed is of the order of about 4 percent depending on water depth. This small variation along with the sound speed minima has a decided effect on propagation of sound over long ranges.

Sound speed minima are axes of sound channels where rays are trapped or concentrate, whereas the maxima cause a divergence among rays. The sound-trapping region near the surface is called the surface duct, whereas the sound-trapping region at depth is called the deep sound channel or SOFAR (Sound Fixing and Ranging) channel. Figure 5-13 shows the latitudinal behavior of the deep sound channel axis depth in the North Atlantic. The deep sound channels are deepest in midlatitudes. An examination of the temperature vs. depth plots of figure 5-6 indicates that this is caused by: (1) very steep thermoclines and shallow mixed layers of tropical regions, and (2) near uniformity with depth of temperature and salinity caused by convective mixing in high latitudes.

e. Modes of Sound Propagation in the Ocean

The typical sound speed profile of deep oceans results in four major modes of sound propagation, depending on the depths of the sound source and receiver. In figure 5-14A, the surface duct traps rays by successive bounces off the surface and upward bending by the increasing sound speed with depth. In figure 5-14B, rays from a source at the deep sound channel axis are trapped by continued refraction due to the positive gradients on either side of the channel axis. In figure 5-14C, the convergence zone mode returns a ray bundle by total refraction to the surface from great depths. Figure 5-14D shows the bottom and surface bounce mode that prevails for steep rays, relatively shallow water depths, or high surface temperatures.

6. Waves in the Ocean

a. The Sea Surface and Acoustic Effects

The state of the sea surface has important consequences for ocean acoustics. When wind blows over the water, there is a transfer of energy from wind to water that sets the sea surface in motion. Surface waves are generated and a surface drift current is created.

About 20 percent of the water energy goes into surface drift currents and the rest into surface waves. Ordinary wind-driven waves are classified as capillary waves if surface tension controls the formation, or gravity waves if gravity controls wave structure.

The shape of deepwater waves depends to large extent on their length. Gravity waves that are longer than about 1.73 cm are controlled by gravitational and inertial forces. Wave shapes range from a nearly sinusoidal form to a sharply peaked crest and rounded trough which can be approximated by a trochoid (fig. 6-1). A trochoid is the curve generated by a point on the radius of a circle as the circle rolls uniformly on a line. Water particles execute small circular orbits in response to the passing wave, the size of the orbits diminishing with increasing depth below surface (fig. 6-2). Surface waves shorter than 1.73 cm in length are generally controlled by forces associated with surface tension, causing the short capillary waves to have rounded crests and v-shaped troughs. Figure 6-3 shows the shape of capillary waves in terms of the amplitude-to-wavelength parameter (a/λ).

Ocean waves are dispersive, since wave speed depends on wavelength or frequency (fig. 6-4). Gravity waves show normal dispersion, since the wave speed increases with increase in wavelength. Capillary waves show anomalous dispersion, since their wave speed increases with decreasing wavelength. The minimum wave speed for combined gravity and capillary waves is 23 cm/s, corresponding to a wavelength of 1.73 cm. The wave frequency ω , is related to wave number k , for gravity waves through the expression: $\omega^2 = gk$ where g is the gravitational constant. For capillary waves the relationship is: $\omega^2 = \tau k^3 / \rho$, where: ρ is the water density and τ is the air-water surface tension coefficient, equal to 73 dyn/cm at 20°C.

Wind effects on the sea surface are not simple, because turbulence and viscosity in both sea and air introduce complex nonlinear effects. Wind turbulence creates a moving pattern of minute fluctuations in air pressure over water, which can generate the initial tiny ripples that eventually become fully developed waves. It is thought that much of the grip that wind has on water to produce waves and currents is provided through those very small wavelets, since they are very numerous and move slowly before the wind. Gravity waves speed up with growth to the point where they can eventually move at the same speed as the wind and can extract no further energy from it. Swell is the name given to those gravity waves that move away from the generating area faster than the wind.

Ordinary gravity waves normally observed at sea are composed of two varieties: swell, which is a long and relatively symmetrical wave, having a period in the order of 10 seconds, produced by wind and storms at some distance from the point of observation, and sea, which is the local wind-generated wave motion, usually of shorter period, with unsymmetrical slopes and steep or white-capped crests depending upon the wind speed and fetch. Fetch is the length of sea surface over which the wind blows without appreciable change of direction.

For practical purposes, a sea-state scale based on visual estimates relating wave height, wind speed, and sea surface appearance has been quite useful. One, based on the Naval Oceanographic Office code, is given in figure 6-5.

An important application of the sea-state code to underwater sound is the Knudsen curves for average ambient noise in water shown in figure 6-6. The curves fall off at the rate of about 5 dB/octave.

Another application of the sea-state code in terms of crest-to-trough wave height occurs in the propagation of sound involving ray paths that bounce off the sea surface. Propagation loss per bounce (a_3) is given as a function of the product of mean wave height, $h(\text{ft})$, and acoustic frequency, $f(\text{kHz})$.

$$a_3 = 1.64 (fh)^{1/2} \text{ dB/bounce}$$

This relation is plotted in figure 6-7.

b. Energy Spectrum of the Wind Waves

Actual waves on the sea surface are irregular, aperiodic, and short crested. The wave spectrum concept may be used to describe distribution of energy among waves of different frequency or wave number. Wave spectrum must be studied in terms of probabilistic models and measured and analyzed by statistical techniques. A wave model is used to describe fluctuation in wave height at any point in terms of a statistically invariant gaussian function of time.

As the wind blows and as waves grow, turbulent variations and varying amounts of internal eddy viscosity -- as well as the interaction of one wave with another -- set an individual limit to the growth of each individual component wave train in the developing sea. They do this by initiating the energy-dissipating, breaking, or "white-cap" process, whenever a momentarily high crest reaches an unstable configuration. In classical theory, this occurs when the height-to-length ratio of waves in a train reaches the critical value of 1/7 at which point in a wave's development its crest is sharply peaked. Thus, the total energy present in all waves on a developing sea progressively distributes itself over a range of frequencies, each frequency characterizing a particular wave train. As waves continue to grow and as new trains continually develop, this range extends more and more to lower frequency -- or longer period -- waves. In brief, a spectrum of ocean waves is formed (fig. 6-8), in which -- for any given wind velocity and for fully developed waves -- energy distribution over a narrow band of wave frequencies from 0.05 to 0.3 Hz is distinctive.

The accepted expression for the energy spectrum of wind-driven gravity waves is that of Pierson and Moskowitz (PM). It is a directional spectrum and is given by:

$$S(\omega, \theta, \nu) = (ag^2/\omega^5) [\exp -\beta (\omega_0/\omega)^4] F(\omega, \theta^*, \nu) m^2s$$

where: $a = 8.1 \times 10^{-3} \text{ m}^2/\text{s}$, $\beta = 0.74$ and $\omega_0 = g/\nu_{19.5}$ rad/sec

$\nu_{19.5}$ is the wind speed at 19.5 meters above the sea surface, and $F(\omega, \theta^*, \nu)$ is the beamwidth of the wave spectra depending on wave angular frequency (ω), wind speed (ν), and direction (θ^*) measured with respect to direction of the wind and is between $\pm \pi/2$. For the fully risen or equilibrium sea, this spectrum falls off as ω^{-5} . The exponential term gives growth portion of the curve.

c. The Thermocline and Internal Waves

Fluctuations, variations, and waves in the temperature structure of the thermocline occur quite frequently in the ocean and can have important consequences for duct propagation. The sharp density gradients of the thermocline give rise to gravity oscillations, called internal waves (fig. 6-9). Oceanic internal waves generally have wave-lengths, periods, and amplitudes much larger than surface waves. Internal waves travel much slower than surface waves. These waves may have periods measured in minutes and hours, wavelengths in kilometers, and amplitudes of the order of tens of meters.

There are theoretical maximum and minimum frequencies for free internal waves. The theoretical maximum frequency is known as the Vaisala frequency. The theoretical lower limit is defined by the Coriolis parameter. The Vaisala frequency N is expressed in terms of density ρ , density gradient $d\rho/dz$, gravity g , and sound speed C , in the medium as follows:

$$N^2 = g \left(\frac{1}{\rho} \frac{d\rho}{dz} - \frac{g}{C^2} \right)$$

The maximum frequency is a function of depth; it is largest in the thermocline and least at depth of weakest density gradient. A continuous monotonically decreasing power spectrum seems to exist between minimum and maximum frequencies along with line frequencies of internal waves.

Several different types of forces have been identified as causing internal waves. Tidal and other forces that impel water movement around land boundaries and topographic features can start oscillations in the thermal structure. Strong winds may create convection cells and eddies in the upper layers of the sea, and the resulting circulation will cause current shears and waves on the thermocline. Vertical oscillations in the thermocline can be induced by air pressure changes and fluctuating winds at the sea surface. Currents crossing sills or underwater mountain ranges may lead to internal waves.

Power spectra of internal waves seem to show regions of distinct frequency behavior (fig. 6-10). Power spectra show that the greatest

power density is in the low-frequency region. The power spectra decrease monotonically with increase in frequency until a frequency of about 0.02 cpm, at which point they fall off less rapidly with frequency at about $f^{-5/3}$ (k^{-1}). At a frequency of about 0.8 cpm there appears to be another transition zone corresponding to decay at a rate of $f^{-5/2}$ (k^{-2}).

d. Acoustic Propagation and Internal Waves

Horizontal thermoclines show a shadow zone beyond the limiting ray in the duct. Within the duct, more or less acoustic energy is trapped depending upon whether the limiting ray is steeper or shallower than with an undisturbed thermocline. The effect of internal waves on acoustic propagation is to form regions of convergence and divergence of rays below the thermocline. In particular, figure 6-11 demonstrates the effect on propagation in and below a surface duct. There is an increase of energy leakage out of the duct bounded by a wavy thermocline. In addition, there are alternating regions of high and low intensity below it.

FIGURE 2-1. THE INTERIOR LAYERS OF THE EARTH. (ADAPTED FROM M.G. GROSS, OCEANOGRAPHY, CHARLES E. MERRILL, COLUMBUS, OHIO, 1967.)

LAYERS OF THE EARTH AND THEIR DENSITIES

LAYER	DEPTH RANGE km	DENSITY g/cm ³ APPROXIMATE AVERAGE
CRUST	0-33	2.8
MANTLE	33-2900	4.5
OUTER CORE	2900-5100	11.4
INNER CORE	5100-6371	12.2
TOTAL EARTH	0-6371	5.5

FIGURE 2-2. EARTH'S CRUST CONSISTS OF UNITS MOVING AS RIGID BLOCKS. BOUNDARIES BETWEEN BLOCKS ARE RIDGES (RIGES), TRENCHES (OR FOLDED MOUNTAIN BELTS) AND FAULTS. (ADAPTED FROM M.G. GROSS, OCEANOGRAPHY, SECOND EDITION, CHARLES E. MERRILL PUBLISHING COMPANY, 1971, PAGE 24.)

FIGURE 2-3. FORMATION OF OCEANIC CRUST AT RIDGES AND DESTRUCTION IN TRENCHES (SCHEMATIC REPRESENTATION). (ADAPTED FROM M.G. GROSS, OCEANOGRAPHY, SECOND EDITION, CHARLES E. MERRILL PUBLISHING COMPANY, 1971, PAGE 21.)

FIGURE 2-4. BOTTOM PROFILE OF THE NORTH ATLANTIC OCEAN (CAPE HENRY, VIRGINIA, TO RIO DE ORO, AFRICA). (ADAPTED FROM B.C. HEEZEN, M. THARP AND M. EWING, THE GEOLOGICAL SOCIETY OF AMERICA, 1959, PL. 22.)

FIGURE 2-5. BOTTOM TOPOGRAPHY CHART OF THE NORTH ATLANTIC OCEAN.
(NAVAL OCEANOGRAPHIC OFFICE, 1966)

FIGURE 2-6. THE HYPSOGRAPHIC CURVE.

FIGURE 2-7. DISTRIBUTION OF DEEP-SEA SEDIMENTS. THE MOUTHS OF THE TEN LARGEST RIVERS ARE SHOWN. (MODIFIED AFTER G. ARRHENIUS, "PELAGIC SEDIMENTS", IN THE SEA: IDEAS AND OBSERVATIONS ON PROGRESS IN THE STUDY OF THE SEAS, VOL. 3, ED. M. N. HILL [NEW YORK: INTERSCIENCE PUBLISHERS, 1963], P. 657.)

FIGURE 2-8. TYPICAL METHOD OF CONTINUOUS SEISMIC PROFILING, NORMAL INCIDENCE REFLECTION ABOVE AND OBLIQUE REFLECTION BELOW, USED TO MEASURE SUB-BOTTOM STRUCTURE. (ADAPTED FROM M. N. HILL, ED., *THE SEA*, VOL. 3, INTERSCIENCE, NEW YORK, 1963.)

FIGURE 2-9. BOTTOM LOSS FOR "FAST" AND "SLOW" SEDIMENTS—RAYLEIGH MODEL
(ACOUSTIC FREQUENCY = 1 kHz).

FIGURE 3-1. THE CLASSIFICATION OF THE BENTHIC AND PELAGIC MARINE ENVIRONMENT.
 (AFTER J.W. HEDGPETH, ED., *TREATISE ON MARINE ECOLOGY AND PALEOECOLOGY*,
 ECOLOGY, VOL. 1. THE GEOLOGICAL SOCIETY OF AMERICA, BOULDER,
 COLORADO, 1957.)

FIGURE 3-2. TRANSPARENCY CHART OF THE NORTH ATLANTIC OCEAN.
(ADAPTED FROM M.A. FREDERICK, U.S. NAVAL POSTGRADUATE SCHOOL, 1970.)

FIGURE 3-3. A PLANKTON SAMPLE WITH REPRESENTATIVES OF THE MEROPLANKTON.
(NAVAL OCEANOGRAPHIC OFFICE)

ZOOPLANKTON

MEROPLANKTON

FIGURE 3-4. REPRESENTATION OF THE DEEP SCATTERING LAYER (DSL). (ECHOSOUNDER RECORD $f = 12 \text{ kHz}$) (SINGLE LOCATION) (NAVAL OCEANOGRAPHIC OFFICE, 1966).

FIGURE 3-5. VARIATION OF EFFECTIVE SCATTERING AREA WITH WAVELENGTH FOR SMALL SOLID PARTICLES AND BUBBLES. (NAVAL OCEANOGRAPHIC OFFICE, 1966)

FIGURE 3-6. REPRESENTATIVE FISHES POSSESSING SWIMBLADDERS.
(OCEANOGRAPHER OF THE NAVY, OCEAN SCIENCE PROGRAM OF THE
U.S. NAVY, 1970.)

LANTERNFISH
MYCTOPHIDAE)

1 INCH

FIGURE 4-1. DISSOLVED CHEMICAL CONSTITUENTS OF SEAWATER (BASED ON CHLORINITY = 19‰.)
(SVERDRUP, JOHNSON, AND FLEMING, *THE OCEANS*, © 1942, RENEWED 1970,
PRENTICE-HALL, ENGLEWOOD CLIFFS, N. J.).

MATERIAL	CONCENTRATION, g/1 kg SEAWATER	ELEMENT/cl‰	PERCENT SALT BY WEIGHT
CHLORIDE, Cl^-	18.980	0.99894	55.04
SODIUM, Na^+	10.556	0.5556	30.61
SULFATE, SO_4^{2-}	2.649	0.1394	7.68
MAGNESIUM, Mg^{2+}	1.272	0.06695	3.69
CALCIUM, Ca^{2+}	0.400	0.02106	1.16
POTASSIUM, K^+	0.380	0.02000	1.10
BICARBONATE, HCO_3^-	0.140	0.00735	0.41
BROMIDE, Br^-	0.065	0.00340	0.19
BORIC ACID, H_3BO_3	0.026	0.00137	0.07
STRONTIUM, Sr^{2+}	0.013	0.00070	0.04
FLUORIDE, F^-	0.001	0.00007	0.00
TOTAL	34.482‰		99.99%

FIGURE 4-2. DEFINITION OF SALINITY.

"SALINITY IS THE TOTAL AMOUNT IN GRAMS OF SOLID MATERIAL DISSOLVED IN 1kg OF SEAWATER WHEN ALL CARBONATE HAS BEEN CONVERTED TO OXIDE. ALL THE IODINE AND BROMINE HAVE BEEN REPLACED BY CHLORINE, AND ALL ORGANIC MATTER HAS BEEN COMPLETELY OXIDIZED."

IT IS EXPRESSED AS GRAMS OF SALT PER KILOGRAM OF SEAWATER.

THE FOLLOWING EMPIRICAL RELATIONSHIP BETWEEN SALINITY AND CHLORIDE CONTENT (CHLORINITY) IS USED TO CONVERT A MEASUREMENT OF CHLORIDE CONTENT TO SALINITY.

$$S^{\text{‰}} = 1.80655 \text{ Cl}^{\text{‰}}$$

FIGURE 4-3. GASES IN DEEP OCEAN WATER AS COMPARED WITH THEIR CONCENTRATION IN AIR. (MODIFIED FROM R. REVELLE AND H.E. SUESS, THE SEA, VOL. 1, INTERSCIENCE, N.Y., 1963, PAGE 313.)

	MILLILITER PER LITER AIR	MILLILITER PER LITER WATER	PERCENTAGE OF TOTAL GAS ON EARTH DISSOLVED IN OCEAN
O ₂	210	2 to 8	0.5
N ₂	781	13	0.6
He	52×10^{-4}	0.5×10^{-4}	0.3
Ne	182×10^{-4}	1.8×10^{-4}	0.3
A	9.32	0.32	1.2
Kr	10×10^{-4}	0.6×10^{-4}	2
Xe	0.8×10^{-4}	0.07×10^{-4}	3
CO ₂	0.3	50	98

FIGURE 4-4. ABSORPTION OF WAVES IN SEAWATER.

FIGURE 4-5. SPEED OF SOUND IN SEAWATER (WILSON)

$$C = 1449.14 + C_t + C_p + C_s + C_{stp}$$

$$C_t = 4.5721t - 4.4532 \times 10^{-2}t^2 - 2.6045 \times 10^{-4}t^3$$

$$+ 7.9851 \times 10^{-6}t^4$$

$$C_p = 1.60272 \times 10^{-1}p + 1.0268 \times 10^{-5}p^2$$

$$+ 3.5216 \times 10^{-9}p^3 - 3.3603 \times 10^{-12}p^4$$

$$C_s = 1.39799 (S-35) + 1.69202 \times 10^{-3} (S-35)^2$$

$$C_{stp} = (S-35) (-1.1244 \times 10^{-2}t + 7.7711 \times 10^{-7}t^2$$

$$+ 7.7016 \times 10^{-5}p - 1.2943 \times 10^{-7}p^2 + 3.1580 \times 10^{-8}pt$$

$$+ 1.5790 \times 10^{-9}pt^2) + p (-1.8607 \times 10^{-4}t$$

$$+ 7.4812 \times 10^{-6}t^2 + 4.5283 \times 10^{-8}t^3)$$

$$+ p^2 (-2.5294 \times 10^{-7}t + 1.8563 \times 10^{-9}t^2)$$

$$+ p^3 (-1.9646 \times 10^{-10}t).$$

WHERE: C IS SOUND SPEED IN m/s

t IS TEMPERATURE IN °C ($-4^{\circ}\text{C} \leq T \leq 30^{\circ}\text{C}$)

S IS SALINITY IN ‰ ($0 \text{ ‰} \leq S \leq 37 \text{ ‰}$)

p IS PRESSURE IN kg/cm² OR ATMOSPHERES

($1 \text{ kg/cm}^2 \leq P \leq 1000 \text{ kg/cm}^2$)

J. ACOUST. SOC. AM. 32, 1357 (1960).

FIGURE 4-6. MAGNESIUM SULPHATE COMPLEX IN SOLUTION.

FIGURE 4-7. ABSORPTION OF SOUND IN SEAWATER (SCHULKIN AND MARSH, 1963).

$$\alpha = \left(\frac{2.34 \times 10^{-6} S f_t f^2}{f_t^2 + f^2} \right) + \left(\frac{3.38 \times 10^{-6} f^2}{f_t} \right) \left(1 - 6.54 \times 10^{-4} p \right) \text{ NEPERS/METER}$$

- WHERE: f_t IS $21.9 \times 10 \left(6 - \frac{1520}{t + 273} \right)$ kHz
 t IS TEMPERATURE IN °C
 f IS FREQUENCY IN kHz; $f \geq 2.5$ kHz
 p IS PRESSURE IN kg/cm² OR ATMOSPHERES
 S IS SALINITY IN ‰
 α IS THE ABSORPTION IN NEPERS/METER

FIGURE 4-8. ABSORPTION OF SOUND IN SEAWATER (THORP, 1967).

LOW-FREQUENCY ANOMALY

$\alpha = \frac{0.1f^2}{1+f^2} +$

$\text{MgSO}_4 \text{ RELAXATION}$

$\frac{40f^2}{4100+f^2} +$

VISCOUS ABSORPTION

$0.000275f^2$

α = ATTENUATION IN dB/kyd

f = FREQUENCY IN kHz

FIGURE 5-1. RADIATION BALANCE IN THE NORTHERN HEMISPHERE AND AVERAGE OCEAN SURFACE TEMPERATURES. (ADAPTED FROM M. G. GROSS, OCEANOGRAPHY, SECOND EDITION, CHARLES E. MERRILL PUBLISHING COMPANY, 1971, PAGE 73.)

FIGURE 5-2. A SCHEMATIC REPRESENTATION OF THE CELLULAR CIRCULATION OF THE ATMOSPHERE ON A ROTATING EARTH. THE DRIVING FORCE FOR THIS MOTION IS THE THERMAL IMBALANCE BETWEEN POLE AND EQUATOR. THE DIVISION OF THE CIRCULATION PATTERN INTO THREE CELLS BETWEEN POLE AND EQUATOR RATHER THAN A SINGLE CELL IS CAUSED BY THE EARTH'S ROTATION. (AFTER EARTH SCIENCE CURRICULUM PROJECT, INVESTIGATING THE EARTH, HOUGHTON MIFFLIN, BOSTON, 1967.)

FIGURE 5-3. WATER MOVEMENTS IN A WIND-GENERATED CURRENT IN THE NORTHERN HEMISPHERE. (M.G. GROSS, OCEANOGRAPHY, SECOND EDITION, CHARLES E. MERRILL PUBLISHING COMPANY, 1971, PAGE 91.)

FIGURE 5-4. GENERALIZED ATMOSPHERIC AND OCEANIC CIRCULATION. (R. H. FLEMING, 1957)

NE: NORTH EQUATORIAL CURRENT
 EC: EQUATORIAL COUNTER CURRENT
 SE: SOUTH EQUATORIAL CURRENT

FIGURE 5-5. SIMPLE THREE-LAYERED OCEAN IN WINTER
SHOWING RELATIVE VOLUME IN EACH ZONE
(NATIONAL DEFENSE RESEARCH COMMITTEE
V. 6A, 1946).

FIGURE 5-6. REPRESENTATIVE WATER TEMPERATURE PROFILES BY SEASON AND LATITUDE.
(NAVAL OCEANOGRAPHIC OFFICE, 1966)

FIGURE 5-7. WIND AND CONVECTIVE MIXING.

FIGURE 5-8. PERCENT OF OCCURRENCE OF LAYER DEPTH IN NORTH ATLANTIC (BY SEASON)

SEASON	LAYER DEPTH (ft)			
	≤ 50	> 50 TO 250	> 250 TO 450	> 450
WINTER (DEC TO MAR)	0	2	33	65
SPRING (APR TO MAY)	1	54	12	33
SUMMER (JUN TO AUG)	5	95	0	0
FALL (SEPT TO NOV)	0	90	10	0

FIGURE 5-9. THE CHANGE IN AVERAGE MID-OCEAN SURFACE SALINITIES IN RESPONSE TO LATITUDINAL CHANGES IN EVAPORATION AND PRECIPITATION. (BASED ON DATA OF H. U. SVERDRUP, FROM A. N. STRAHLER PHYSICAL GEOGRAPHY, JOHN WILEY, 1960).

FIGURE 5-10. TEMPERATURE AND SALINITY OF THE WATER MASSES OF THE ATLANTIC OCEAN, AS DERIVED FROM METEOR STATIONS 74, 88, AND 246. (AFTER SVERDRUP, JOHNSON, AND FLEMING, THE OCEANS, © 1942, RENEWED, 1970, PRENTICE-HALL, ENGLEWOOD CLIFFS, N. J.)

FIGURE 5-11. THE GEOGRAPHIC DISTRIBUTION OF WATER MASSES THAT HAVE COMMON VERTICAL SALINITY AND TEMPERATURE DISTRIBUTIONS, AS DENOTED BY THEIR T—S CURVES. SQUARES MARK THE REGIONS IN WHICH THE CENTRAL WATER MASSES ARE FORMED; CROSSES INDICATE THE LINES ALONG WHICH THE ANTARCTIC AND ARCTIC INTERMEDIATE WATERS SINK. (SVERDRUP, JOHNSON, AND FLEMING, THE OCEANS, © 1942, RENEWED 1970, PRENTICE-HALL, ENGLEWOOD CLIFFS, N.J.)

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FIGURE 5-12. BASIC TEMPERATURE AND SOUND VELOCITY STRUCTURE OF THE OCEAN.

FIGURE 5-13. SOUND CHANNEL AXIS DEPTH VS LATITUDE (55° W LONGITUDE)

FIGURE 5-14. SOUND TRANSMISSION MODES
(NAVAL OCEANOGRAPHIC OFFICE, 1966).

FIGURE 6-1. SIMPLE SEA WAVE AND ITS PARTS. (M.G.GROSS,
OCEANOGRAPHY, SECOND EDITION, CHARLES E. MERRILL,
1971.)

FIGURE 6-2. WAVE PROFILE AND WATER-PARTICLE MOTIONS CAUSED BY A WAVE IN DEEP WATER. NOTE THE DIMINISHING SIZE OF THE ORBITS WITH INCREASING DEPTH BELOW THE SURFACE. (M.G. GROSS, OCEANOGRAPHY, SECOND EDITION, CHARLES E. MERRILL, 1971.)

LITTLE WAVE MOTION BELOW THIS DEPTH

FIGURE 6-3. STREAMLINES OF THE FLOW IN FINITE AMPLITUDE CAPILLARY WAVES. ANY MEMBER OF THE FAMILY SPECIFIES THE SHAPE OF THE FREE SURFACE FOR A CERTAIN VALUE OF α/λ ; THE LOWER CURVES THEN REPRESENT INTERNAL STREAMLINES. (G.D. CRAPPER, 1957)

FIGURE 6-4. THEORETICAL SPEED OF GRAVITY WAVES AND CAPILLARY WAVES AS A FUNCTION OF THEIR WAVELENGTH.

FIGURE 6-5. SEA STATE CODE (WORLD METEOROLOGICAL ORGANIZATION
(NAVAL OCEANOGRAPHIC OFFICE, 1966))

SEA STATE	WAVE HEIGHT RANGE METERS	WIND SPEED RANGE m/s	TERM AND EFFECTS OBSERVED
0	0	0.0-1.5	SEA LIKE MIRROR. RIPPLES WITH APPEARANCE OF SCALES; NO FOAM CRESTS.
1	0 < 0.3	1.6-3.3	SMALL WAVELETS; CRESTS OF GLASSY APPEARANCE, NOT BREAKING.
2	0.3 < 0.6	3.4-5.4	LARGE WAVELETS; CRESTS BEGIN TO BREAK; SCATTERED WHITECAPS.
3	0.6 < 1.2	5.5- 7.9	SMALL WAVES, BECOMING LONGER; NUMEROUS WHITECAPS.
4	1.2 < 2.4	8.0-10.7	MODERATE WAVES TAKING LONGER FORM; MANY WHITECAPS, SOME SPRAY.
5	2.4 < 4.0	10.8-13.8	LARGER WAVES FORMING; WHITECAPS EVERYWHERE; MORE SPRAY.
6	4.0 < 6.1	13.9-24.4	SEA HEAPS UP, WHITE FOAM FROM BREAKING WAVES BLOWN IN STREAKS. HIGH WAVES; SEA BEGINS TO ROLL; DENSE STREAKS OF FOAM.
7	6.1 < 9.1	24.5-28.4	VERY HIGH WAVES WITH OVERHANGING CRESTS; SEA WITH VERY DENSE FOAM STREAKS; VISIBILITY REDUCED.
8	9.1 < 13.7	28.5-32.6	EXCEPTIONALLY HIGH WAVES; SEA COVERED WITH FOAM; VISIBILITY MORE REDUCED.
9	>13.7	32.7-61.2	AIR FILLED WITH FOAM; SEA COMPLETELY WHITE WITH DRIVING SPRAY; VISIBILITY GREATLY REDUCED.

WAVE HEIGHT IS DEFINED AS THE MEAN TROUGH TO CREST VALUE.

FIGURE 6-6. DEEPWATER AMBIENT NOISE LEVELS (KNUDSEN CURVES) FOR SEA STATES 0 TO 6.
(NAVAL OCEANOGRAPHIC OFFICE, 1966)

FIGURE 6-7. SEA STATE AND ACOUSTIC LOSS AT THE SEA SURFACE.

FIGURE 6-8. COMPARISON OF AN OBSERVED WAVE SPECTRUM TO EMPIRICAL SPECTRA OF PIERSON-MOSKOWITZ. (PM).

FIGURE 6-9. ISOTHERMS IN INTERNAL WAVE MOTIONS. THE UPPER BLOCK SHOWS A DOMINANT COMPONENT WITH TIDAL FREQUENCY, AND AMPLITUDE ABOUT 15 m (J. F. REID, 1956). THE LOWER BLOCK ILLUSTRATES THE OSCILLATIONS IN A SHALLOW THERMOCLINE WITH A CONSIDERABLY HIGHER FREQUENCY (E. C. LA FOND, 1962).

FIGURE 6-10. INTERNAL WAVE SPECTRA MEASURED BY CHARNOCK (1965). THE MEASUREMENTS WERE MADE BY TOWING A THERMISTOR CHAIN RAPIDLY THROUGH THE WAVE FIELD ON VARIOUS COURSES, SO THAT THE FREQUENCIES OBSERVED ARE THE FREQUENCIES OF ENCOUNTER AS THE CHAIN TRAVERSES THE FIELD. THE DIFFERENT SYMBOLS REPRESENT RESULTS FROM SECTIONS TAKEN IN DIFFERENT DIRECTIONS. THE LACK OF ANY SYSTEMATIC VARIATION WITH COURSE INDICATES THE ABSENCE OF STRONG DIRECTIONALITY IN THE INTERNAL WAVES. THE LOWER SCALE GIVES EQUIVALENT WAVE-NUMBER TIMES THE MEAN THERMOCLINE DEPTH. (H. CHARNOCK, 1965).

FIGURE 6-11. DIAGRAM OF RAYS IN THE SOUND MEDIUM, WHICH HAS AN INTERNAL WAVE ON THE THERMOCLINE. (LEE, 1961)

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20. ABSTRACT (Continue on reverse side if necessary and identify by block number) A brief introduction is given of oceanography for acousticians. The main topics and related acoustical examples are: submarine geology (bottom acoustic loss); marine biology (the deep sound scattering layer); chemical and physical properties of seawater (sound speed and sound absorption); the steady state of the ocean (sound speed vs depth structure and modes of sound propagation in the ocean); and waves in the ocean (sea surface loss and sea surface noise, and acoustic propagation and internal waves).		

19.

waves in the ocean

sea surface loss

sea surface noise

acoustic propagation

internal waves