U.S. DEPARTMENT OF COMMERCE
ENVIRONMENTAL SCIENCE SERVICES ADMINISTRATION

DEEP CIRCULATION, CENTRAL
NORTH PACIFIC OCEAN:
1961, 1962, 1963

WILLIAM D. BARBEE

COAST AND GEODETIC SURVEY
RESEARCH PAPER (1965)

MBL/WHOI

0 0301 0001832 1

U.S. DEPARTMENT OF COMMERCE

John T. Connor, *Secretary*

ENVIRONMENTAL SCIENCE SERVICES ADMINISTRATION

Robert M. White, *Administrator*

COAST AND GEODETIC SURVEY

James C. Tison, Jr., *Director*

Research Paper

Deep Circulation, Central
North Pacific Ocean:
1961, 1962, 1963

By

William D. Barbee
Commander, USESSA

UNITED STATES GOVERNMENT PRINTING OFFICE: WASHINGTON; 1965

For sale by the Superintendent of Documents, U.S. Government Printing Office Washington, D.C., 20402—Price \$1.75

Preface

Findings presented in this paper are based upon field observations collected as part of the Coast and Geodetic Survey's oceanographic survey program of the central North Pacific Ocean. The paper was prepared originally as a thesis in partial fulfillment of the requirements for the degree of Master of Science in oceanography at the University of Washington, Seattle, Wash., in 1964. As a thesis, the paper was titled "Deep Circulation in a Portion of the North Pacific Ocean During the Summers of 1961, 1962, and 1963." The author, who participated in the North Pacific Ocean survey during 1961, 1962, and 1963, is grateful for the opportunity afforded him by the Coast and Geodetic Survey to pursue graduate study in oceanography at the University of Washington. He would like to acknowledge the interest and support shown by the faculty and staff of the Department of Oceanography at the university during preparation of the thesis—particularly Dr. Clifford A. Barnes for helpful guidance, Dr. Maurice Rattray, Jr., for critical review, Mrs. Monique Rona for machine processing of data, and Mr. Donald R. Doyle for preparation of illustrations. Machine processing of data and preparation of drawings were supported in part by the U.S. Office of Naval Research (Contract Nonr 477-10, Project NR 083012). The paper is published in its original form except for editorial revision of format. Illustrations are grouped at the end of the paper. The list of symbols is in appendix III.

Contents

	<i>Page</i>
Preface.....	iii
Abstract.....	1
1. Introduction.....	1
1.1 Objectives.....	1
1.2 Observations.....	2
1.3 Description of the area.....	3
2. Statement of the problem.....	5
2.1 Outline of procedure.....	5
2.2 Assumptions and approximations.....	6
3. Results.....	10
3.1 Horizontal velocities.....	10
3.1.1 Velocities in 1961.....	10
3.1.2 Velocities in 1962.....	10
3.1.3 Velocities in 1963.....	11
3.2 Vertical velocity.....	12
3.3 Depth of no horizontal motion.....	12
3.4 Annual cycle of horizontal velocities.....	12
4. Summary and conclusions.....	15
References.....	17
Appendix I. Physical expressions for the problem.....	19
Appendix II. A numerical example.....	27
Appendix III. List of symbols.....	35
Figures 1 through 68.....	37-104

Tables

TABLE	<i>Page</i>
1. Recent Coast and Geodetic Survey observations in the central North Pacific Ocean.....	3
2. Data printout for section defined by oceanographic stations 63 and 59.....	31
3. Data printout for section defined by oceanographic stations 29 and 32.....	32
4. Data printout for section defined by oceanographic stations 63 and 29.....	33
5. Data printout for section defined by oceanographic stations 59 and 32.....	34

Illustrations

FIGURE	<i>Page</i>
1. Locations of stations and sections.....	37
2. Stations used in determination of transport.....	38
3. Temperature (°C) Section 1, September 1961.....	39
4. Salinity (‰) Section 1, September 1961.....	40
5. Dissolved oxygen (ml/L) Section 1, September 1961.....	41
6. Temperature (°C) Section 2, September 1961.....	42
7. Salinity (‰) Section 2, September 1961.....	43
8. Dissolved oxygen (ml/L) Section 2, September 1961.....	44
9. Temperature (°C) Section 3, September 1962.....	45
10. Salinity (‰) Section 3, September 1962.....	46
11. Dissolved oxygen (ml/L) Section 3, September 1962.....	47

FIGURE	Page
12. Temperature ($^{\circ}\text{C}$) Section 4, September 1963.....	48
13. Salinity ($^{\circ}/_{\text{oo}}$) Section 4, September 1963.....	49
14. Dissolved oxygen (ml/L) Section 4, September 1963.....	50
15. Temperature ($^{\circ}\text{C}$) Section 5, May 1963.....	51
16. Salinity ($^{\circ}/_{\text{oo}}$) Section 5, May 1963.....	52
17. Dissolved oxygen (ml/L) Section 5, May 1963.....	53
18. Variation with latitude of wind-driven integrated geostrophic transport, September 1962.....	54
19. Geopotential anomaly at 4,000 meters referred to 5,000 meters, Sections 2, 3, and 5.....	55
20. Geopotential anomaly at 4,000 meters referred to 5,000 meters, Sections 1 and 4.....	56
21. Geopotential anomaly at 3,000 meters referred to 5,000 meters, Sections 2, 3, and 5.....	57
22. Geopotential anomaly at 3,000 meters referred to 5,000 meters, Sections 1 and 4.....	58
23. Topography in dynamic meters, 250-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	59
24. Topography in dynamic meters, 500-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	60
25. Topography in dynamic meters, 1,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	61
26. Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	62
27. Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	63
28. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	64
29. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	65
30. Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, August 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	66
31. Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, August 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	67
32. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, August 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.....	68
33. Topography in dynamic meters, 250-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.....	69

FIGURE	Page
34. Topography in dynamic meters, 500-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.....	70
35. Topography in dynamic meters, 1,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.....	71
36. Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.....	72
37. Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.....	73
38. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.....	74
39. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, September 1962. Drag coefficient 0.0026; meaned stresses from twice-daily reports of pressure.....	75
40. Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, September 1962. Drag coefficient 0.0012; meaned stresses from twice-daily reports of pressure.....	76
41. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, September 1962. Drag coefficient 0.0012; stresses meaned from twice-daily reports of pressure.....	77
42. Topography in dynamic meters, 250-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.....	78
43. Topography in dynamic meters, 500-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.....	79
44. Topography in dynamic meters, 1,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.....	80
45. Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.....	81
46. Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.....	82
47. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.....	83
48. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.....	84
49. Topography in dynamic meters, 250-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; meaned stresses from twice-daily reports of pressure.....	85
50. Topography in dynamic meters, 500-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; meaned stresses from twice-daily reports of pressure.....	86

FIGURE	Page
51. Topography in dynamic meters, 1,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.....	87
52. Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.....	88
53. Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.....	89
54. Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.....	90
55. Variation of vertical velocity with depth at indicated latitudes along approximately 170° W.....	91
56. Depth of no zonal motion, 160° W, 1961.....	92
57. Depth of no zonal motion, 177° W, 1961.....	93
58. Depth of no meridional motion, 169° W, 1961.....	94
59. Depth of no zonal motion, 160° W, 1962.....	95
60. Depth of no zonal motion, 165° W, May 1963.....	96
61. Depth of no zonal motion, 180° W, May 1963.....	97
62. Depth of no meridional motion, 172° W, May 1963.....	98
63. Annual variations in zonal velocity at 4,000 m.....	99
64. Variation with latitude of integrated total transport, September 1961.....	100
65. Variation with latitude of zonal Ekman transport, September 1961.....	101
66. Variation with latitude of zonal Ekman transport, September 1961.....	102
67. Variation with longitude of meridional Ekman transport, September 1961.....	103
68. Variation with latitude of integrated geostrophic transport, September 1961...	104

Deep Circulation, Central North Pacific Ocean: 1961, 1962, 1963

Abstract. Methods are reviewed for obtaining velocity reference values such that the absolute flow in the deep sea can be described in terms of these reference values and the dynamic height computations. The method of Murty and Rattray (1962) for obtaining absolute flow by adjusting the transports from dynamic height computations to analogous values of transport determined from the curl of the wind stress was examined and used for calculating velocities for a portion of the central North Pacific Ocean between the Hawaiian and Aleutian Islands from observations made during the summers of 1961, 1962, and 1963.

A predominantly eastward flow was found at depths of 250 and 500 meters. This flow is similar to the accepted pattern of flow in this area. Over much of the area velocities at 4,000 and 5,000 meters were on the order of 1 cm/sec and in a direction opposite to those near the surface. The flow near the bottom tends to conform to salient features of the bathymetry. By comparison with flow inferred from the distribution of temperature and dissolved oxygen, it is concluded that the flow at depths of 4,000 meters and more computed for the summer months does not reflect the net annual circulation, but is a periodic barotropic flow.

Errors in the assumptions in the formulation of the equations used for computing the flow, in the inadequacies in tabulations of wind-driven transport, and in observational errors are examined.

1. Introduction

1.1 Objectives

The method of computing relative velocities by the dynamic heights method has become standard in oceanography. Dynamic height computations can give a realistic representation of the absolute velocity field in respect to the earth if proper allowances can be made for wind drift and if the relative velocities can be referred to absolute velocities at some level. However, the estimation of absolute velocities for reference has often lacked rigor.

The main objective of this paper is to describe as realistically as possible, the circulation in a portion of the North Pacific Ocean. A method is developed which relates transports calculated from oceanographic station data to geostrophic transports determined from the wind-stress field to find values of the velocity near the bottom. These velocities near the bottom are then used as references for the dynamic height velocities. The method is applied using three recent series of oceanographic observations and tabulations of wind-driven geostrophic transports from Fofonoff (1960).

Defant (1961) reviewed methods of defining and delineating a reference surface for dynamic topography. He cited the method of determining absolute from relative values by comparing direct current measurements with the relative velocities. However, he pointed out that the difficulty in making a sufficient number of direct measurements, each long enough to define the mean field of motion, drastically limits the use of this method.

In practice, a reference surface is selected arbitrarily near the ocean bottom or as deep as the observations allow. Surface and near-surface flow generally can be realistically estimated because velocities at the reference depth are usually negligible in comparison to velocities near the surface. However, this arbitrary selection of a reference level can result in relatively large errors in computing velocities of the deeper currents,

and in computations of transport. Generally, the depth of zero velocity will not be a level surface, but will vary from place to place. In some early attempts to define this variable-depth reference surface, it was placed at the depth of minimum dissolved oxygen. However, it is difficult to accept the assumption that the minimum of dissolved oxygen is a consequence of no horizontal advection. Further, velocities obtained using this reference surface were unrealistic.

Defant calculated the depth of the reference surface by a method which is implicit in the dynamic height observations. He used his method to construct a chart of the depth of the reference surface in the Atlantic Ocean. A plot was made of pressure (or depth) versus the difference in dynamic heights for each pair of adjacent stations. The plotted curve represents the velocity component normal to the section. Any part of the plotted curve that is parallel to the pressure coordinate represents a depth interval of constant velocity. Defant reasoned that the velocity is more likely to be zero in this layer of no shear stress than at any other level. In constructing his chart of the depth of this reference surface, he made the additional assumption that the horizontal flow parallel to the dynamic height section is also zero at this level. Although this method of selecting a reference surface is subjective, its use has resulted in consistently realistic velocity fields. This is essentially the method used by Russian investigators working with 1958–1959 *Vityaz* data from the North Pacific Ocean (Chekotillo, 1961).

Stommel (1956) described a method for determining the depth of no meridional motion by equating the divergence of the Ekman transport with the divergence of the geostrophic transport. Using the concept introduced by Fofonoff (1962) that the total transport can be separated into a barotropic mode, a baroclinic mode, and an Ekman transport, Stommel's method can be expressed as a relationship between these three modes:

$$\rho_m v_b = \frac{1}{h} (V - V_E - V_g)$$

where the V 's represent mass transport per unit width, v_b is the velocity at the bottom, h is the total depth, ρ_m is the mean density of the water column, and the subscripts E and g denote Ekman and baroclinic modes, respectively. Stommel did not consider the effects of bottom topography on the divergence of the barotropic mode of transport.

Murty and Rattray (1962), Fofonoff (1961), Favorite (1961), and Dodimead, Favorite, and Hirano (1962) have compared baroclinic transports with wind-driven transports. The method used by Murty and Rattray, in which the baroclinic transport is compared with the wind-driven geostrophic transport to obtain a velocity near the bottom, is used herein. The method is developed in appendix I and a numerical example is given in appendix II.

Russian workers have determined deep currents by similar methods. Koshlyakov (1961) reported computations in the North Pacific Ocean based on arguments similar to Stommel's, but included the effects of variations in depth by means of an expression similar to:

$$\rho_m v_b = \frac{\beta(V_E + V_g) - \left(\frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y} \right)}{\frac{\partial(fh)}{\partial y}}$$

In this expression the new terms τ_{sx} and τ_{sy} are wind stresses at the surface, β is the rate of change of the Coriolis parameter with latitude, and f is the Coriolis parameter. This expression is valid in areas where the product of the zonal component of velocity at the bottom and the zonal component of the bottom slope is much smaller than the product of the meridional components of the same variables.

1.2 Observations

In 1961, the Coast and Geodetic Survey of the United States Department of Commerce began field work on a survey of that part of the North Pacific Ocean between the Hawaiian Islands on the south, the Aleutian Islands on the north, and (approximately) the 158° W and 180° meridians on the east and west, respectively. Oceanographic stations in this program included observations of temperature by reversing thermometers, salinity by salinometer, and dissolved oxygen by Winkler titration (table 1 and fig. 1). Only those stations deeper than 2,000 meters (fig. 2) were useful for the present study. The field observations are summarized graphically in figures 3 through 17.

TABLE 1.—*Recent Coast and Geodetic Survey observations in the central North Pacific Ocean*

Year	Dates	Vessel	Sections	Number of deep stations	National Oceanographic Data Center reference number
1961	10 Sept.–12 Oct.	<i>Pioneer</i>	160 °W. long.	6	31099
			177 °W. long.	6	31099
1962	22 Aug.–25 Sept.	<i>Pioneer</i>	160 °W. long.	10	31099
1963	6–29 May	<i>Pioneer</i>	165 °W. long.	10	
	6–29 May	<i>Surveyor</i>	180 ° long.	6	

Values for the variables at standard depths and computed values of sigma- t , the specific volume anomaly, the geopotential anomaly, and the potential energy anomaly (Fofonoff, 1962), were determined by means of an interpolation and computation program prepared by the Applied Mathematics staff of the Department of Oceanography, University of Washington. The computations were made on the University's IBM 709 computer. The 1961 and 1962 observations were also submitted to the National Oceanographic Data Center, where values for standard depths were again calculated. The values reported by the National Oceanographic Data Center disagree slightly with those from the University because of minor variations in the two interpolation and computation programs.

1.3 Description of the Area

The observations listed above are of considerable interest in supporting recent descriptions of the fields of oceanographic variables in the central North Pacific Ocean. Since 1955, co-operative efforts such as those summarized by the NORPAC Committee (1960a, b), observations resulting from International Geophysical Year cruises, programs of the International North Pacific Fisheries Commission and of the Bureau of Commercial Fisheries in connection with its Pacific Oceanic Fisheries Investigations, have provided a basis for detailed descriptions of the area by previous investigators. The present investigation confirms these descriptions of temperature, salinity and oxygen distributions (figs. 3 to 17 inclusive), and amplifies certain details. However, the primary purpose of this investigation is to describe the circulation and characteristics of the deep water.

Most oceanographic descriptions of this area treat the upper 1,000 meters because that stratum includes the more striking parts of the structure, and relatively few data are available from greater depths. Hence, most descriptions of the circulation deal with surface velocities or are limited to velocities and transports above 1,000 meters. Reid (1961) described the surface geostrophic flow for the Pacific Ocean. His calculations were based on a level of no net motion at 1,000 decibars. The NORPAC Atlas (1960a) covers in greater detail the geostrophic flow in the North Pacific Ocean. Permanent features of the surface flow in the central north Pacific Ocean are: the *Alaskan Stream* and

Alaskan Stream Extension, which form a relatively intense westerly flow south of and closely confined to the Aleutian Island Arc. A zone of easterly flow lies between 30° N and 50° N. Dodimead, Favorite, and Hirano (in press) designate parts of this easterly flow as the *Subarctic Current*, made up in large part of Oyashio waters; the *Westwind Drift*, originating where parts of the Oyashio and Kuroshio meet and turn eastward; and the *North Pacific Current*, formed by the main flow of the Kuroshio which turns easterly at about 35° N. From 30° N to the southern limits of the present study area the flow is weak, as would be expected near the center of the anticyclonic flow in the subtropical north Pacific Ocean. A westerly flow persists at the southern limit ($23^{\circ}30'N$) of the present investigation.

Strong gradients of temperature and salinity are found across the Westwind Drift. This band of strong gradients, extending nearly across the north Pacific Ocean is called the *Polar Front*. Fleming (1955, 1958), Uda (1963), and Dodimead, Favorite and Hirano (in press) divide the North Pacific Ocean into natural regions based on the current systems and the associated characteristics of temperature, salinity and dissolved oxygen. The southern edge of the Polar Front is the boundary between the *Subarctic Region* in the north and the *Subtropic Region* in the south. Dodimead and coworkers define the boundary between the two regions as the zone where the 34‰ isohaline is nearly vertical from the surface to a depth of 300 to 500 meters (figs. 4, 7, 10, 13, and 16). In the present study area, the boundary occurs at about 40° N.

Above 1,000 meters, the density structure in the Subarctic Region is dominated by salinity effects. Seasonal variations in the salinity structure are exhibited only in the zone extending from the surface to about 100 meters. Below this zone a permanent halocline, in which the salinity increases rapidly, occurs in the 100- to 200-meter depth interval. Below the halocline, the salinity gradually increases with depth. Fleming (1958), and Tully and Barber (1960) have described the salinity structure in detail. The negligible seasonal variation in all but the upper 100 meters of the water column is explicit in their descriptions. The temperature and salinity structures in the Subarctic Region are subject to pronounced seasonal variations through the same depth interval. A temperature minimum is often present in the depth interval of the halocline as shown in the northern end of the sections (figs. 6, 9, and 12). The salinity distribution dominates the density structure and maintains stability despite the temperature inversion.

The halocline becomes progressively less pronounced southward through the Polar Front, and finally disappears.

In the Subtropic Region there is no permanent halocline. Here the density structure in the upper 1,000 meters is dominated by a strong thermocline extending to depths of 600 to 800 meters (figs. 3, 6, 9, 12, and 15). A salinity minimum occurs at depths of 500 to 600 meters (figs. 4, 7, 10, 13, and 16). Seasonal effects in the surface layers in the Subtropic Region are less pronounced than in the Subarctic Region.

The division of the North Pacific Ocean into regions on the basis of the temperature and salinity distributions applies between the surface and 1,000 meters. Below 1,000 meters, temperature and salinity structures are similar throughout the area.

The dissolved oxygen structure is characterized by a minimum occurring between 600 and 1,200 meters. It occurs at about 600 meters immediately south of the Aleutian Island Arc, descends to about 1,200 meters in the region of the Polar Front, and ascends through the Subtropic Region to about 800 meters near $23^{\circ}30'N$. A noteworthy feature of the meridional sections of oxygen concentration (figs. 5, 8, 11, 14, and 17) is the thickening of the zone of low oxygen values in the northern part of the area. From $46^{\circ}N$ to $52^{\circ}N$, the thickness of the oxygen zone, of concentrations less than 1 milliliter/liter, is twice that south of $40^{\circ}N$.

Figures 19 through 54 show the results of the dynamic computations for the sections investigated in this study. These will be considered in the text which follows.

2. Statement of the Problem

2.1 Outline of Procedure

The basic premise is that the mass transport calculations tabulated by Fofonoff (1960) describe the total transport, and that they can be used as reference values for transport calculations by the dynamic heights method. The procedure for determining these reference values is as follows:

- a. The geostrophic mode of the total transport is extracted from Fofonoff's tabulation for the section separating two oceanographic stations.
- b. Transport through the same section is computed from the dynamic heights at the two stations. These dynamic height calculations are based on a level at 4,000 or 5,000 meters, whichever more nearly equals the mean depth of the section.
- c. The transport determined in b (as described by the dynamic heights) is the total *baroclinic transport* (provided the flow due to internal structure vanishes at the bottom) and the difference between the transports determined by methods a and b must be the *barotropic transport*. This barotropic transport divided by the water density and the area of the section gives the barotropic velocity.
- d. Combining the barotropic velocity with baroclinic velocities determined from the dynamic height calculations results in a description of the total geostrophic flow through the section. If the assumption in c (that baroclinic flow is zero at the bottom) is not valid, then the division of the flow into barotropic and baroclinic modes is indeterminate; but the description of the total geostrophic flow and its variation with depth is still valid.

To establish a basis in theory for the method, to determine the limitations of (and the assumptions made in) Fofonoff's (1960) transport calculations, and to justify using his results in the present manner, it is necessary to develop applicable expressions for transport. This has been done in appendix I. Based on Fofonoff's (1962) treatment of Munk's (1950) wind-driven flow formulation, the expressions developed for barotropic velocities u_b and v_b are:

$$\begin{aligned}
 u_b &= \frac{(\psi_{\phi b} - \psi_{\phi a}) - \sum_{\phi a}^{\phi b} U_E R \Delta\phi - \left(\frac{\chi_{\phi b} - \chi_{\phi a}}{f} \right)}{\rho_m h R \Delta\phi} \\
 v_b &= \frac{(\psi_{\lambda b} - \psi_{\lambda a}) - \sum_{\lambda a}^{\lambda b} V_E R \cos \phi \Delta\lambda - \left(\frac{\chi_{\lambda b} - \chi_{\lambda a}}{f} \right)}{\rho_m h R \cos \phi \Delta\lambda}
 \end{aligned} \tag{35a, b}$$

In (35a), ψ is a transport function for total transport, and so the first term in the numerator on the right expresses the total transport through a meridional section. Zonal Ekman transport per unit length along the meridian is denoted U_E , R is the radius of the earth, and $\Delta\phi$ is a change in latitude, so the second term is Ekman transport through the same section. The final term is baroclinic transport, expressed as the difference in *potential energy anomalies*, χ , across the section, divided by the Coriolis parameter, f . The denominator on the right side is the mean density, ρ_m , times the area of the section, $hR\Delta\phi$.

2.2 Assumptions and Approximations

The validity of results from the approach used here is dependent on the assumptions (a) that the wind-driven geostrophic transport tabulated by Fofonoff (1960) is accurate in the region of interest; (b) that acceleration terms, bottom friction terms and horizontal friction terms can be neglected in developing the expressions for wind-driven transport (p. 21, appendix I); and (c) that convergence due to the effect of bottom topography can be neglected. The results will also depend on (d) the accuracy, precision, number and distribution of the oceanographic observations.

(a) Fofonoff (1960, p. 3), in considering the applicability of his computations of wind stress at the surface, stated, "It should be remembered that the transports are very sensitive to the proportionality factors used in relating geostrophic to surface wind and surface wind to wind stress. Numerical equivalence of computed and observed transports is not anticipated." His calculations of surface stress are based on the equation

$$\tau_s = \rho_a C_D w^2,$$

where τ_s is the surface stress, ρ_a is the air density, C_D is the drag coefficient, and w is the wind speed. It is assumed that the wind speed and stress magnitude are functions of the pressure difference, Δp , between positions so that

$$\tau_s \propto w^2 \propto (\Delta p)^2.$$

Additional error is introduced in Fofonoff's computed stresses by using the square of monthly means of pressure differences, $(\Delta p)^2$, rather than computing the mean stress from the actual pressure (that is, satisfying the relationship $\tau_s \propto \Delta p^2$). The error arising from calculation of stresses from a monthly mean of pressure differences rather than using the monthly mean of stresses estimated from continuous reports of pressure is proportional to the variance of the pressure differences. Since Δp^2 is larger than $(\Delta p)^2$, the less rigorous relationship leads to stresses that are too small.

Fofonoff (1960) used a value of 0.0026 for the drag coefficient in his tabulation of wind-driven transport. Deacon and Webb (1962) have summarized recent determinations and find a range of values from about 0.0010 to 0.0015. Because of the linear relationships between the drag coefficient and Ekman and total transport, this discrepancy increases Fofonoff's values by a factor of about 2.

Graphic comparisons have been made to assess the effects of these two discrepancies. Figure 18 shows the alternate determinations for the variation with latitude of integrated geostrophic transport in September 1962 (Section 3). The three determinations are based on:

- (i) stresses obtained by using a drag coefficient of 0.0026 and monthly means of pressure differences;
- (ii) a drag coefficient of 0.0026 and a monthly mean of values estimated from twice-daily pressure reports; and,
- (iii) a drag coefficient of 0.0012 and a monthly mean of the values estimated from twice-daily pressure reports.

The use of the higher drag coefficient (of 0.0026) rather than the more nearly correct value of 0.0012 leads to high transport values; but the use of monthly mean pressures (rather than means from twice-daily values) gives low transports. Accordingly, errors in method (i) are partly compensated and the results obtained agreed well with method (iii); whereas method (ii) gives erroneously high results in all cases. Figures 38, 39 and 41 show for the 5,000 meter level the dynamic topography resulting from the September 1962, observations adjusted to these three techniques of wind-driven transport computation.

The effects of adjusting the dynamic topography to the geostrophic transport by methods (i) and (iii) can be seen by comparing figures 42 to 47 (drag coefficient of 0.0026 and monthly mean pressures) with figures 49 to 54 (drag coefficient 0.0012 and twice-daily mean pressures). These figures show the same general pattern and similar slopes with discrepancies of only 1 to 2 dynamic centimeters in a few locations.

(b) The calculations for wind-driven transport for each month are based on the stress-distribution for that month's mean pressure distribution, without consideration of initial conditions or the inertial response characteristics of the ocean. For the assumption to hold that acceleration terms are negligible, the ocean system must be essentially in equilibrium with wind-imposed forces during each month considered. Intuitively, a response time shorter than one month would seem realistic for the barotropic mode, but intuition as well as analysis of observations precludes such short response times for the baroclinic mode. Veronis and Stommel (1956), in their treatment of variable wind stress on a two-layered ocean without lateral boundaries, showed that for periods of the order of one month, acceleration terms were unimportant in both the barotropic and baroclinic modes. An explanation for this is that the barotropic mode has a response time much shorter than one month and so it is at all times essentially in equilibrium with the driving force, while the baroclinic mode primarily responds to forces with periods of the order of one year or more. Presumably fluctuations of wind stress for periods shorter than one month will give insignificant accelerations.

In addition to wind stress, bottom and lateral friction must also be considered. Proudman (1953) calculated bottom stress using the square law

$$\tau_b = c_b \rho_b u_b^2$$

in which the drag coefficient, c_b , is taken as 0.0025, ρ_b , the density of water at the bottom is roughly 1 gm/cm³, and u_b , the velocity in cm/sec, is measured one meter above the bottom. In the present investigation, deep velocities on the order of 1 cm/sec were found. Under these conditions, bottom stresses would be of the order of 0.002 dynes/cm² and negligible compared to surface stresses.

Munk (1950) developed expressions for wind-driven transport in a central region of the ocean in which horizontal friction terms were neglected. These expressions result in a distribution of transport conforming to the distribution of properties, and thus support the assumption that horizontal friction is negligible in central oceanic regions.

(c) The assumption that divergence due to the effect of bottom topography is negligible in the area of study is examined here. Sverdrup (1947), Munk (1950), and Fofonoff (1962) all specified that the transport was made up of only baroclinic and Ekman modes; therefore, the treatment of the bottom convergence problem was not pertinent. If equations (22), (23), and (24) from appendix I are added and the term for bottom effects retained:

$$\beta(V_g + V_E + V_b) = \frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y} - \rho' f \left(u_b \frac{\partial z_b}{\partial x} + v_b \frac{\partial z_b}{\partial y} \right).$$

The term for bottom effects can be expressed as

$$\rho' f (|\vec{v}_b| \cos \theta |\nabla z_b|),$$

where θ is the angle between the bottom velocity and gradient vectors. Substituting and rearranging,

$$V_b = \frac{\left(\frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y} \right)}{\beta} - V_g - V_E - \frac{\rho' f (|\vec{v}_b| \cos \theta |\nabla z_b|)}{\beta}.$$

The meridional barotropic transport can also be approximately expressed as

$$V_b \simeq h v_b \rho_m.$$

As noted in the discussion of the accuracy and precision of observations in section (d) which follows, velocities are significant to within about 0.6 cm/sec. Representative values of the depth, h , and the mean density, ρ_m , are 500,000 cm and 1 gm/cm³, respectively, so the resulting difference in barotropic transport for a velocity difference of 0.6 cm/sec would be

$$V_b \sim 3 \cdot 10^5 \text{ gm cm}^{-1} \text{ sec}^{-1},$$

and terms on the right side of the expression for barotropic transport of order $10^4 \text{ gm cm}^{-1} \text{ sec}^{-1}$ and less could be neglected. Total transport and barotropic transport terms are of the order $10^5 \text{ gm cm}^{-1} \text{ sec}^{-1}$. The Ekman transport is of the order 10^4 or $10^5 \text{ gm cm}^{-1} \text{ sec}^{-1}$.

In examining the data from September 1961 (fig. 28), it is apparent that the most critical combination of bottom velocity and bottom slope occurs at the Hawaiian Island Rise. Between stations 25 and 26, the bottom velocity is about 2 cm/sec. The bottom slope of $2 \cdot 10^{-3}$ was computed from the smoothed isobaths on H.O. Chart 5486 and is in general agreement with that computed from station soundings. The angle θ which the stream lines make with a line normal to the isobaths is uncertain because of the lack of detail of bottom topography and the general character of the velocity determinations. It is apparent, however, that the bottom water tends to flow along the bottom contours, a tendency particularly evident in the northern and southwestern portions of the area observed (figs 27 and 28). Assuming a value of θ of 85° (suggested in the present data by the close approach of the direction of the bottom currents to the isobaths), bottom velocity as 2 cm/sec, $\frac{f}{\beta}$ as 3×10^8 for 25° N , and $\rho = 1 \text{ gm/cm}^3$, the magnitude of the final term in the expression for barotropic transport is approximately $10^5 \text{ gm cm}^{-1} \text{ sec}^{-1}$. This magnitude in a depth of 5,000 meters results in a constant velocity of 0.2 cm/sec—one third the significant value of 0.6 cm/sec (see below).

(d) Since the ratios of the baroclinic to barotropic modes within the flow are computed from oceanographic observations, the accuracy and the precision of these observations must be examined for their influence on the velocity at any level. Velocity at the bottom is of primary interest. If values of $\pm 0.01^\circ \text{C}$ and $\pm 0.01\text{‰}$ are accepted as the precision of temperature and salinity measurements and the difference between stations is such as to give the maximum error in specific volume anomaly for the interval of 0 to 1,000 decibars, an error of ± 0.01 dynamic meter will arise. Wooster and Taft (1958), using a statistical approach to this problem, found that the standard error of the difference in geopotential anomaly (0 over 1,000 db) between two stations is 0.0056 dynamic meter and conclude that the measurement error of this difference is ± 0.011 dynamic meter (26). For a pair of stations at 30° latitude separated by 250 km, the error in the computed geostrophic velocity is $\pm 0.6 \text{ cm/sec}$. These values of anomaly and velocity must be kept in mind when assessing the significance of the horizontal velocities determined in this program. The method used, equating baroclinic transport plus barotropic transport to wind-driven geostrophic transport, partially compensates for errors in the observations. Any observational error leading to a baroclinic transport too large will result in barotropic transport too small by the same amount. Thus the barotropic velocity, constant with depth, will be slightly too small. The error in total computed velocity at the depth of the observational error will be slightly less than that in the baroclinic velocity and almost negligible at other depths.

The significant value of geopotential anomaly (0.01 dynamic meter) must also be considered in assessing the significance of computed vertical velocities. Since the flow determined here is geostrophic, the variation of vertical velocity, w , with depth can be approached through the expression

$$\frac{\partial w}{\partial z} = \frac{\beta}{f} v,$$

where v is the meridional component of total velocity and β and f are as previously defined. In the present study area the horizontal flow is mainly zonal; meridional velocities are usually of the same order as the level of significance determined for horizontal velocities.

So the values of $\frac{\partial w}{\partial z}$ obtained by the above expression are questionable.

To help establish the consistency of the observations, and also to determine the validity of the proposition that baroclinic velocity vanishes at the bottom, plots were made of the geopotential anomaly as a function of latitude. Figures 19 and 20 show the geopotential anomaly at 4,000 meters referred to 5,000 meters; figures 21 and 22 are of the anomaly at 3,000 meters referred to 5,000 meters. At 4,000 meters, the greatest range of the anomaly is 0.02 dynamic meter, only twice the measurement error. This does not permit the assumption that baroclinic flow in the interval from 5,000 to 4,000 meters can be ignored, but it does indicate the relative unimportance of this deep internal structure. The geopotential anomaly at 3,000 meters referred to 5,000 meters varies more with time and place than does the anomaly at 4,000 meters. Effects of water structure in the 3,000- to 4,000-meter interval must certainly be included in the determinations.

3. Results

3.1 Horizontal Velocities

Dynamic heights were calculated and adjusted to give transports equal to those determined from the curl of the wind stress (see appendix II for a numerical example). These adjusted heights are shown as lines of equal geopotential through the area of the investigation (figs. 23 through 54). The selected contour interval is 0.01 dynamic meter, a value which is uncertain in the upper levels where the paucity of stations must result in excessive smoothing of the dynamic topography. But this interval is used throughout to facilitate comparison of all levels.

3.1.1 Velocities in 1961.—In figures 23 through 28, the wind-driven geostrophic transport for September 1961 was used to obtain reference values for dynamic height calculations based on oceanographic observations made during that month. Geopotential anomalies at the 250, 500, 1,000, 2,000, 4,000, and 5,000 meter levels are presented. The geostrophic transport for these figures was determined from the difference between wind-driven total transport and Ekman transport. Figure 29 shows the geopotential anomalies at 5,000 meters calculated from the September 1961 oceanographic data and the wind-driven geostrophic transport which was computed from the tabulated geostrophic transport function. Figures 30 through 32 show the results of a comparison of the flow computed from the geostrophic transport function in August 1961.

The flow pattern was generally zonal, especially in the upper layers. North of about 40° N, the direction of flow was easterly at 250, 500, and 1,000 meters, but reversed to westerly at 2,000, 4,000, and 5,000 meters. South of 30° N, a significant southerly flow occurred at 250 and 500 meters. At 2,000 meters, and below, the flow conformed to the trend of the bottom contours along the Hawaiian Island Rise. The deep velocities were westerly north of the rise and easterly south of it. Figures 28 and 29 can be used to compare the consequences of using the wind-driven flow with and without consideration of Ekman divergence to obtain reference values for dynamic heights. No significant differences are noted. An analysis of the oceanographic observations used in these determinations (or of other observations in the same area) shows the water structure leading to baroclinic transport to vary only slightly. A comparison between September 1961 oceanographic data adjusted for transports from the wind stresses of September (figs. 26, 27, and 28) and then of August (figs. 30, 31, and 32) gives an indication of the variations in the barotropic velocity.

The flow at 4,000 and at 5,000 meters tended to conform to the isobaths. This tendency was evident both at the Hawaiian Island Rise and near the Aleutian Trench.

3.1.2 Velocities in 1962.—The oceanographic observations in 1962 comprise a single section along the 160°W meridian. By adjusting the values of the dynamic heights to yield transports equal to the geostrophic transports arising from different wind stress distributions, an evaluation is made of the effects of (a) estimating wind stress from a monthly mean of atmospheric pressure distributions contrasted to estimating it from twice-daily reports of pressure distribution and taking a monthly mean of these values, and (b) using different drag coefficients in combination with the two alternate methods of finding the mean monthly stress values. Figures 33 through 38 show the geopotential topography adjusted to geostrophic transport arising from wind stresses based on a drag coefficient of 0.0026 and the mean pressure distribution for September 1961. Figure 39 is based on the same drag coefficient but the average of two daily reports of atmospheric pressure

were used in computing stresses. Figures 40 and 41 are based on a drag coefficient of 0.0012 and twice-daily reports of atmospheric pressure.

The flow patterns as computed vary with the method of calculating stresses and particularly marked differences in these computed patterns occur in high latitudes. Comparison of figure 38 with figure 39 shows that alternative methods of computing mean stresses give apparent velocities in opposite directions north of 45°N. However, when both the drag coefficient and the method of obtaining mean stresses were changed (figs. 40 and 41), the resulting velocities at equal levels were but little different up to 45°N from those indicated in figures 36 and 38. It should be noted that Ekman divergence is not considered in figures 36 and 38, and its significance is discussed in a following section. The use of 0.0026 as the drag coefficient resulted in values for stress that apparently were too large, but calculating stress values from a monthly mean of pressure distribution rather than estimating stresses from twice-daily reports of pressure and then taking a monthly mean led to stress values that apparently were too small, and these discrepancies tend to cancel one another.

3.1.3 Velocities in 1963.—The oceanographic data for 1963 are presented in the same manner as the 1961 data. Figures 42 through 47 are of geopotential topography adjusted to wind-driven geostrophic transport with consideration of Ekman divergence. The wind-driven transport is based on a drag coefficient of 0.0026, and the mean pressure distribution for May 1963 is used. As in the 1961 results, flow was mainly zonal. North of 40° N, the flow at 250, 500, and 1,000 meters was easterly. Velocities decreased from about 3 cm/sec at 250 meters to less than 1 cm/sec at 1,000 meters. From 40° to 30° N, flow was southerly. At 500 meters, flow was easterly or northeasterly north of 30° N; it was southwesterly south of 30° N. At the 1,000-meter level, the flow south of 40° N was poorly defined in the 1963 data, though there were indications of a cyclonic flow centered at about 35° N, 180°. South of 28° N, there was a weak westerly set. At 2,000 meters, flow was weak over most of the area and there is some indication of an anticyclonic circulation centered near 35° N, 180°. The patterns of the flow at 4,000 and 5,000 meters were similar to each other. North of 45° N, velocities at these depths were of the order of 1 cm/sec westerly, and were the largest found at or below 4,000 meters in 1963. Between 45° and 40° N, the flow was easterly, shifting to southerly south of 40° N.

Flow at the deep levels tended to conform to the salient features of the bottom topography. This tendency—more obvious at 5,000 meters (fig. 47) than at 4,000 meters (fig. 46)—was exhibited near the bathymetric feature at 30°N, 162°W, near the Aleutian Trench, and in the vicinity of the Hawaiian Island Rise. Throughout much of the area, circulation at 4,000 and 5,000 meters was opposite to that above 1,000 meters.

Figure 48 depicts the dynamic topography adjusted to the May 1963 wind-driven geostrophic transport in which the Ekman divergence is neglected. Differences between the flow patterns shown in figure 48, which neglects Ekman divergence, and figure 47, which includes consideration of Ekman divergence, were not significant.

At the northern limit of the area, near the Aleutian Islands, the dynamic topography suggests a boundary current in a direction opposing the flow at lower latitudes. In this region of high lateral friction, details of current patterns are open to some question as the method of computing wind-driven transport used is an approximation applicable to central ocean areas.

Figures 49 through 54 depict the dynamic topography adjusted to wind-driven geostrophic transport which was determined from the May 1963 mean of stresses calculated from twice-daily reports of pressure, using a drag coefficient of 0.0012. Results were similar to those shown in figures 42 through 47 (based on the higher drag coefficient and an alternative method of estimating mean values as discussed previously). The more accurate determination of wind-driven transport suggests a slightly weaker flow at the 1,000 meter level south of 40° N.

3.2. Vertical Velocity

The rate of change of vertical velocity with depth, $\frac{\partial w}{\partial z}$, was determined from the adjusted values of meridional velocity for May, 1963, and the expression

$$\frac{\partial w}{\partial z} = \frac{\beta}{f} v.$$

With w assumed to be zero at the bottom, $\frac{\partial w}{\partial z}$ was graphically integrated between 5,000 and 250 meters. The distributions of vertical velocity with depth are shown in figure 55. The maximum values determined at the four latitudes investigated varied from $3 \cdot 10^{-5}$ cm/sec to $14 \cdot 10^{-5}$ cm/sec. Though the magnitudes agree with those inferred by Knauss (1962), the directions do not. Knauss inferred upward velocities over the depth interval 2,500 meters to 5,000 meters, whereas a consistent downward direction was found in this study at all levels below 1,000 meters. This disagreement may arise from the fact that Knauss's values reflect the net circulation, while values from the present determination are associated with a specific period. It should also be noted that the values found for $\frac{\partial w}{\partial z}$ are of the same magnitude as the limit of sensitivity accepted for the determination. Uncertainties in the present determinations of vertical velocities and in the relationship of these velocities to net vertical circulation notwithstanding, it is noteworthy that the upward velocity found near 50°N above 1,000 meters is in accordance with flow inferred from oxygen profiles. In figures 5, 11, 14, and 17, the local thickening of the low-oxygen interval and the decrease in depth of oxygen isopleths toward the north in the depth interval 200 to 1,000 meters suggest upward flow in this area.

3.3 Depth of No Horizontal Motion

In the present formulation, a component of the geostrophic velocity is zero at a depth where the barotropic mode is equal and opposite to the baroclinic mode. The depths of no zonal motion or of no meridional motion are shown as functions of latitude in figures 56 through 62.

If, near the depth where the baroclinic and barotropic velocities add up to zero, the baroclinic velocity changes slowly with depth, the determination of the depth of no motion will be sensitive both to observational errors in the oceanographic data and to discrepancies in the values for wind-driven transport. The sensitivity in the determination together with the paucity of data leads to uncertainty in the plots of the surfaces of no zonal or meridional motion. However, figures 56 through 62 do indicate that the depths of no zonal or meridional motion vary widely with time and place, and that flow at depth is frequently opposite to that near the surface.

3.4 Annual Cycle of Horizontal Velocities

Two features of the flow at 4,000 and 5,000 meters persisted through each determination: The well-developed westerly flow north of 45°N, and the southerly flow in the zone from 30° to 35°N (figs. 28, 29, 31, 32, 37, 38, 53, and 54). The southerly flow is in agreement with that inferred by Wooster and Volkmann (1960). Their data are not definitive north of 45°N. However, an easterly velocity can be inferred from the distributions of potential temperature and dissolved oxygen observed in the present study, because potential temperature increases and dissolved oxygen decreases from west to east. So it must be concluded that the velocities determined for September 1961, September 1962, and May 1963 are not representative of the net circulation.

The inferred easterly net flow below 4,000 meters does not preclude the westerly flows determined for specific periods in this depth interval. Each of the determinations in the present study was based on wind-driven geostrophic transports associated with the weak atmospheric pressure gradients of the North Pacific summer (that is, May or September). Thus, total geostrophic transport north of 45° N as found, for example, for the portion of Section 5 between stations 120 and 123 and 125 was small. The mass structure in this horizontal interval is such that a large easterly baroclinic transport existed. The velocities associated with this easterly baroclinic transport are strongest above 2,000 meters, decreasing to weak easterly values below 4,000 meters. The combination of a small geostrophic transport and large easterly baroclinic transport results in a large westerly barotropic transport. Above 2,000 meters, westerly barotropic flow is masked by the stronger easterly baroclinic flow; but below 4,000 meters barotropic flow is stronger than the baroclinic mode, and total flow is westerly. This result is reasonable for the periods associated with the present determinations.

Because the mass structure and the associated baroclinic transport can be expected to be relatively constant, the annual cycle exhibited by the total wind-driven transport must be reflected in the barotropic mode (Fofonoff, 1961c). The baroclinic velocity has been shown to make but a small contribution to the flow below 4,000 meters, therefore a description of the annual cycle in the barotropic flow does much to describe the net flow below 4,000 meters. To gain a qualitative estimate of the annual cycle of the barotropic flow the transport associated with the dynamic heights through stations 120–123 and stations 123–125 (1963) was successively adjusted to monthly values of the wind-driven geostrophic transport. As this facet of the investigation is qualitative at best, it seemed justifiable to use the most readily available tabulation of wind-driven geostrophic transports for the comparison. The 1961 tabulations based on a drag coefficient of 0.0026 and monthly means of the atmospheric pressure distribution were used. Figure 63 is a plot of this determination of the annual cycle of the zonal components of velocities at 4,000 meters. Variations of about 1.5 cm/sec in the zonal velocities are indicated but the net flow is not easterly.

The approach used for estimating the annual cycle of flow shown in figure 63 is deficient in the following respects:

- a. The monthly values used for wind-driven geostrophic transport are only approximations of the proper values. Transport based on a drag coefficient of 0.0012 and mean stresses computed from twice-daily pressure reports might be significantly greater than those used in constructing figure 63, due to the large variance in atmospheric pressure which accompanies the severe, often short-period storms of the North Pacific winter. These larger wind-driven transports should be easterly and larger than the easterly baroclinic transports, thus resulting in easterly geostrophic flow at the 4,000-meter level.

- b. The present formulation is based on the geostrophic approximation that a balance exists between pressure forces and Coriolis forces. It must be recognized that this balance is not perfectly achieved. Those time-dependent motions not satisfying the geostrophic approximation are not considered in the present determinations. These could make a significant contribution to the net flow which has been inferred from the distribution of potential temperature and dissolved oxygen.

- c. The present investigation is based on the assumption that the total wind-driven geostrophic transport and the barotropic transport respond significantly only to disturbances of the one-month period considered. Veronis and Stommel (1956) indicated that the barotropic response might be significant for disturbances with periods of from one week to two months.

- d. In the comparisons supporting figure 63, it is assumed that baroclinic transport is constant and can be accurately described by dynamic height computations based on oceanographic data from May 1963. Obviously, there may be a significant variation in the mass structure and thus in the baroclinic transport through the section.

e. Although the baroclinic mode has been shown to be small in comparison to the barotropic mode below 4,000 meters during three specific periods, it cannot be assumed that the annual mean baroclinic flow is small in comparison to the annual mean barotropic flow. Veronis and Stommel (1956) indicated that net circulation is associated with the baroclinic mode. It can be anticipated that an accurate description of the annual mean baroclinic transport through stations 120–123 and 123–125 might indicate a net easterly flow at the 4,000 meter level.

4. Summary and Conclusions

A method has been derived for matching the geostrophic transport, calculated from dynamic heights, with wind-driven transport for the same period. Geostrophic velocities at 5,000 meters and above were determined covering a portion of the north central Pacific Ocean for the months of September 1961, September 1962, and May 1963. These velocities are reported as plan views of dynamic heights. The determined velocities were generally zonal for the periods of observation. At the 5,000-meter level values on the order of 1 cm/sec were found. Over most of the area at these times, the sense of the velocity changed within the water column, so that the current direction at the 5,000-meter level was opposite to that of the near-surface levels.

The effects of varying the drag coefficient used in determining wind stress and of alternate methods of determining the mean stress distribution for the one-month period of consideration was investigated. The use of a drag coefficient of 0.0012 and values for stresses determined as the mean of stresses calculated from twice-daily reports of pressure distribution seem to give more realistic transports than those obtained from other combinations of drag coefficients and pressures.

The horizontal distribution of potential temperature and dissolved oxygen indicate that the net flow at 4,000 meters at about 48°N must be easterly. Such a flow is in a direction contrary to the flow found at this location by the transport method. This has been accounted for by examining the relative importance of the baroclinic and barotropic modes at this level. The flow at 4,000 meters is mainly barotropic, and as the barotropic flow must be periodic (Veronis and Stommel, 1956) it is concluded that the calculated deep flow is a periodic barotropic flow. The calculated magnitude of deep velocities (about 1 cm/sec) is in good agreement with Veronis and Stommel's theoretical value.

Vertical velocities are determined by graphically integrating $\frac{\partial w}{\partial z}$ determined as a function of meridional velocity. Values as high as 10^{-4} cm/sec were calculated, but these are at the level of significance of observational errors. Thus, the results shown may reflect the range of variation in vertical velocity, but details of the variation with depth cannot be validated.

For each period for which horizontal velocities were computed, plots were made of the surfaces of no zonal or meridional motion. The determination of the depth of no motion can be very sensitive to observational errors and inaccuracies in the computed wind-driven transport; consequently, the plots are subject to considerable error but do indicate that the depth of no zonal or meridional motion varies widely with place and time.

The assumption that terms due to the interaction of flow with the bottom can be ignored was examined. The data suggest that bottom flow parallels isobaths and therefore, the bottom and slope terms in the integrated momentum equations become very small. However, determinations in greater detail over an area in which the bathymetry is well defined would be needed to show conclusively that these terms are negligible.

Because the velocities determined near the bottom appear to be periodic, comparison with net velocities as inferred from the horizontal distribution of variables (Wooster and Volkmann, 1960) or by carbon-14 dating or heat flow considerations (Knauss, 1962) is inappropriate. Perhaps the best corroboration of the deep flow determined in the present study is in the tendency of the water to flow parallel to isobaths.

It is concluded that the method employed here is a useful tool for the study of circulation in central ocean areas. To best realize the possibilities of the method, oceanographic data should be collected so that no additional limitations are placed on the results. A

grid of stations forming sections along meridians and parallels approximately 300 kilometers in length would allow full realization of the capabilities of the method. In addition to a more detailed determination of the flow patterns, such a grid would provide means better to determine the validity of deep observations and to assess the importance of internal structure to the deep flow. Observations in the grid should extend to within a few hundred meters of the bottom.

References

CHEKOTILLO, C. A.

1961. The investigation of horizontal and vertical water circulation in the North Pacific during the winter season, *Abstracts of symposium papers, Tenth Pacific Science Congress, Honolulu, Hawaii*, 1961, pp. 342, 343.

DEACON, E. L. and E. K. WEBB

1962. Interchange of properties between sea and air, small-scale interactions, in M. N. Hill (ed.), *The Sea: Ideas and Observations on Progress in the Study of the Seas*, Vol. 1. Physical Oceanography, Interscience, New York, pp. 43-87.

DEFANT, A.

1961. *Physical Oceanography*, Vol. 1, Pergamon Press, New York. 729 pp.

DODIMEAD, A. J., F. FAVORITE, and T. HIRANO

- In press. *Review of oceanography of the subarctic Pacific region*, Bulletin No. 13, International North Pacific Fisheries Commission, Vancouver, Canada.

FAVORITE, F.

1961. Pacific subarctic circulation, *Abstracts of symposium papers, Tenth Pacific Science Congress, Honolulu, Hawaii*, 1961, pp. 343, 344.

FLEMING, R. H.

1958. Notes concerning the halocline in the Northeastern Pacific Ocean, *J. Mar. Res.*, 17:158-173.

1955. *Review of the oceanography of the North Pacific*, Bulletin No. 2, International North Pacific Fisheries Commission, Vancouver, Canada, 43 pp.

FOFONOFF, N. P.

1960. *Transport computations for the North Pacific Ocean, 1955-1958*, Fish. Res. Bd. Canada, MS Report Series, Oceanogr. and Limnol., Nos. 77-80, 4 volumes.

- 1961a. *Transport computations for the North Pacific Ocean, 1959*, Fish. Res. Bd. Canada, MS Report Series, Oceanogr. and Limnol., No. 85, 181 pp.

- 1961b. *Transport computations for the North Pacific Ocean, 1960*, Fish. Res. Bd. Canada, MS Report Series, Oceanogr. and Limnol., No. 93, 181 pp.

- 1961c. Mass transport in the North Pacific, *Abstracts of symposium papers, Tenth Pacific Science Congress, Honolulu, Hawaii*, 1961, p. 344.

1962. Dynamics of ocean currents, in M. N. Hill (ed.), *The Sea: Ideas and Observations on Progress in the Study of the Seas*, Vol. 1, Physical Oceanography, Interscience, New York, pp. 323-395.

FOFONOFF, N. P., and F. W. DOBSON

- 1963a. *Transport computations for the North Pacific Ocean 1950-1954*, Fish. Res. Bd. Canada, MS Report Series, Oceanogr. and Limnol., Nos. 149-153, 5 volumes.

- 1963b. *Transport computations for the North Pacific Ocean January-May, 1962*, Fish. Res. Bd. Canada, MS Report Series, Oceanogr. and Limnol., No. 164, 35 pp.

FOFONOFF, N. P., and C. K. ROSS

1962. *Transport computation for the North Pacific Ocean 1961.*, Fish. Res. Bd. Canada, MS Report Series, Oceanogr. and Limnol., No. 128, 89 pp.

KNAUSS, J. A.

1962. On some aspects of the deep circulation in the Pacific, *J. Geophys. Res.*, 67(10): 3943-3954.

KOSHLyakOV, M. N.

1961. An example of the computation of the deep currents in the Northeastern Pacific, *Abstracts of symposium papers, Tenth Pacific Science Congress, Honolulu, Hawaii*, 1961, pp. 344, 345.

- MUNK, W. H.
1950. On the wind-driven ocean circulation, *J. Meteorol.*, 7(2):79-93.
- MURTY, C. B., and M. RATTRAY, JR.
1962. On the depth distribution of geostrophic flow in the Eastern North Pacific, *J. Geophys. Res.* 67(4):1649.
- NORPAC Committee
1960a. *Oceanic Observation of the Pacific: 1955, the NORPAC Atlas*, Univ. Calif. Press and Univ. Tokyo Press, Berkeley and Tokyo, 123 plates.
1960b. *Oceanic Observations of the Pacific: 1955, the NORPAC Data*, Univ. Calif. Press and Univ. Tokyo Press, Berkeley and Tokyo, 532 pp.
- PROUDMAN, J.
1953. *Dynamical Oceanography*, John Wiley & Sons, New York, 409 pp.
- REID, J. L.
1961. On geostrophic flow at the surface of the Pacific Ocean with respect to the 1000 decibar surface, *Tellus* 13(4):489-502.
- STOMMEL, H.
1956. On the determination of the depth of no meridional motion, *Deep-Sea Res.* 3(4):273-278.
- SVERDRUP, H. U.
1947. Wind-driven currents in a baroclinic ocean with application to the equatorial currents of the Eastern Pacific, *Proc. Nat. Acad. Sci.*, 33(11):318-326.
- TULLY, J. P., and F. G. BARBER
1960. An estuarine analogy in the sub-arctic Pacific Ocean, *J. Fish. Res. Bd. Canada.*, 17:91-112.
- UDA, M.
1963. Oceanography of the sub-arctic Pacific Ocean, *J. Fish. Res. Bd. Canada*, 20(1): 119-179.
- VERONIS, G., and H. STOMMEL
1956. The action of variable wind stresses on a stratified ocean, *J. Mar. Res.* 15(1): 43-75.
- WOOSTER, W. S., and B. A. TAFT
1958. On the reliability of field measurements of temperature and salinity in the ocean, *J. Mar. Res.* 17:552-566.
- WOOSTER, W. S., and G. K. VOLKMANN
1960. Indications of deep Pacific circulation from the distribution of properties at five kilometers, *J. Geophys. Res.*, 65(4):1239-1249.

Appendix I.

Physical Expressions for the Problem

Sverdrup (1947), Munk (1950), and Fofonoff (1962) have developed expressions relating total transport to the curl of the wind stress. All three formulations were predicated on a velocity structure such that velocity vanished well above the bottom. Thus, it was not necessary to consider the effects of bottom topography on the flow. In the present problem, portions of these existing solutions applicable to central regions are employed. Approximations and assumptions are made pertaining to the effects of bottom topography on the solution for transport.

The development is based on Fofonoff's (1962) treatment of Munk's (1950) wind-driven flow formulation which is in turn, developed from the momentum equations for steady motion. In a coordinate system where x is positive eastward, y is positive northward, and z is positive upward and assuming incompressibility and negligible molecular stress terms, the following equations are obtained:

$$\begin{aligned}\frac{\partial \rho u^2}{\partial x} + \frac{\partial \rho uv}{\partial y} + \frac{\partial \rho uw}{\partial z} - \rho f v &= -\frac{\partial P}{\partial x} + \mu_H \nabla^2 u + \mu_V \frac{\partial^2 u}{\partial z^2} \\ \frac{\partial \rho uv}{\partial x} + \frac{\partial \rho v^2}{\partial y} + \frac{\partial \rho vw}{\partial z} + \rho f u &= -\frac{\partial P}{\partial y} + \mu_H \nabla^2 v + \mu_V \frac{\partial^2 v}{\partial z^2} \\ \frac{\partial P}{\partial z} &= -\rho g\end{aligned}\tag{1a, b, c}$$

and continuity is expressed by

$$\frac{\partial(\rho u)}{\partial x} + \frac{\partial(\rho v)}{\partial y} + \frac{\partial(\rho w)}{\partial z} = 0.\tag{2}$$

Assuming terms with vertical velocities are negligible, the vertically integrated equations of motion can be written as

$$\begin{aligned}\int_b^\eta \frac{\partial \rho u^2}{\partial x} dz + \int_b^\eta \frac{\partial \rho uv}{\partial y} dz - f \int_b^\eta \rho v dz &= - \int_b^\eta \frac{\partial P}{\partial x} dz + A_H \int_b^\eta \rho \left(\frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} \right) dz + \int_b^\eta \rho A_V \frac{\partial^2 u}{\partial z^2} dz \\ \int_b^\eta \frac{\partial \rho uv}{\partial x} dz + \int_b^\eta \frac{\partial \rho v^2}{\partial y} dz + f \int_b^\eta \rho u dz &= - \int_b^\eta \frac{\partial P}{\partial y} dz + A_H \int_b^\eta \rho \left(\frac{\partial^2 v}{\partial x^2} + \frac{\partial^2 v}{\partial y^2} \right) dz + \int_b^\eta \rho A_V \frac{\partial^2 v}{\partial z^2} dz,\end{aligned}$$

where f is the Coriolis parameter, ρ is mean density, and A_H and A_V are the horizontal and vertical kinematic eddy viscosities, respectively.

Convenient terms for the integrals are derived by the following definitions and arguments. First, define the components of horizontal momentum transport

$$\tilde{uv} = \int_{z_b}^\eta \rho uv dz, \quad \tilde{u^2} = \int_{z_b}^\eta \rho u^2 dz;\tag{3a, b}$$

and the components of mass transport, U and V ,

$$\begin{aligned}U &= \int_{z_b}^\eta \rho u dz, \quad V = \int_{z_b}^\eta \rho v dz, \\ \int_b^\eta \frac{\partial P}{\partial x} dz &= \frac{\partial}{\partial x} \int_{z_b}^\eta P dz - P_\eta \frac{\partial \eta}{\partial x} + P_b \frac{\partial z_b}{\partial x},\end{aligned}\tag{4a, b}$$

and, since P_η and $\partial\eta/\partial x$ are both very small,

$$\int_{z_b}^{\eta} \frac{\partial P}{\partial x} dz = \frac{\partial}{\partial x} \int_{z_b}^{\eta} P dz + P_b \frac{\partial z_b}{\partial x}, \quad \int_{z_b}^{\eta} \frac{\partial P}{\partial y} dz = \frac{\partial}{\partial y} \int_{z_b}^{\eta} P dz + P_b \frac{\partial z_b}{\partial y}. \quad (5a, b)$$

Next, define the potential energy of a column

$$E_p = \int_{z_b}^{\eta} P dz \quad (6)$$

so that when equation (6) is substituted in (5a, b)

$$\int_{z_b}^{\eta} \frac{\partial P}{\partial x} dz = \frac{\partial E_p}{\partial x} + P_b \frac{\partial z_b}{\partial x}, \quad \int_{z_b}^{\eta} \frac{\partial P}{\partial y} dz = \frac{\partial E_p}{\partial y} + P_b \frac{\partial z_b}{\partial y}. \quad (7a, b)$$

From equation (1c)

$$\partial z = -\frac{\alpha}{g} \partial P$$

and on substitution into (6)

$$E_p = -\frac{1}{g} \int_{P_b}^0 \alpha P dP. \quad (8)$$

Letting the specific volume, $\alpha = \alpha_0 + \delta$ wherein α_0 is the specific volume of water of salinity 35 $^{\circ}$ / $_{00}$, temperature 0 $^{\circ}$ C, and pressure P ,

$$E_p = \frac{1}{g} \int_0^{P_b} P \alpha_0 dP + \frac{1}{g} \int_0^{P_b} P \delta dP \quad (9)$$

Fofonoff (1962) defines the last term in (9) as the *anomaly of potential energy*,

$$\chi = \frac{1}{g} \int_0^{P_b} P \delta dP. \quad (10)$$

Using the definition in the first term on the left of equation (7a),

$$\frac{\partial E_p}{\partial x} = \frac{\partial \chi}{\partial x} + \frac{\partial}{\partial x} \left[\frac{1}{g} \int_0^{P_b} P \alpha_0 dP \right],$$

and by application of the chain rule

$$\frac{\partial E_p}{\partial x} = \frac{\partial \chi}{\partial x} + \frac{\partial P_b}{\partial x} \frac{\partial}{\partial P_b} \left[\frac{1}{g} \int_0^{P_b} P \alpha_0 dP \right].$$

Since the bracketed portion of the last term is a function of the bottom pressure only,

$$\frac{\partial E_p}{\partial x} = \frac{\partial \chi}{\partial x} + \frac{\partial P_b}{\partial x} \frac{P_b \alpha_0(P_b)}{g}$$

This, and the similar expression for $\frac{\partial E_P}{\partial y}$, substituted in equations (7a, b) give

$$\begin{aligned}\int_{z_b}^{\eta} \frac{\partial P}{\partial x} dz &= \frac{\partial \chi}{\partial x} + \frac{\partial P_b}{\partial x} \frac{P_b \alpha_0(P_b)}{g} + P_b \frac{\partial z_b}{\partial x} \\ \int_{z_b}^{\eta} \frac{\partial P}{\partial y} dz &= \frac{\partial \chi}{\partial y} + \frac{\partial P_b}{\partial y} \frac{P_b \alpha_0(P_b)}{g} + P_b \frac{\partial z_b}{\partial y}.\end{aligned}\quad (11a, b)$$

Finally, it can be shown that the final terms

$$\begin{aligned}\int_{z_b}^{\eta} A_V \frac{\partial^2 u}{\partial z^2} dz &= \tau_{sx} - \tau_{bx}, \\ \int_{z_b}^{\eta} A_V \frac{\partial^2 v}{\partial z^2} dz &= \tau_{sy} - \tau_{by},\end{aligned}\quad (12a, b)$$

where τ_{sx} represents the stress per unit surface area in the x direction and τ_{bx} is a similar term for stress at the bottom. By substituting equations (3a, b), (4a, b), (11a, b), and (12a, b) into (1a, b),

$$\begin{aligned}\frac{\partial \tilde{u}^2}{\partial x} + \frac{\partial \tilde{u}\tilde{v}}{\partial y} - fV &= -\frac{\partial \chi}{\partial x} - \frac{\partial P_b}{\partial x} \frac{P_b \alpha_0(P_b)}{g} - P_b \frac{\partial z_b}{\partial x} + A_H \nabla^2 U + \tau_{sx} - \tau_{bx} \\ \frac{\partial \tilde{u}\tilde{v}}{\partial x} + \frac{\partial \tilde{v}^2}{\partial y} + fU &= -\frac{\partial \chi}{\partial y} - \frac{\partial P_b}{\partial y} \frac{P_b \alpha_0(P_b)}{g} - P_b \frac{\partial z_b}{\partial y} + A_H \nabla^2 V + \tau_{sy} - \tau_{by}.\end{aligned}\quad (13a, b)$$

The bottom pressure and bottom slope terms in (13a, b) can be further simplified as follows:

$$P_b = P_z + g \int_{z_b}^z \rho dz, \quad \frac{\partial P_b}{\partial x} = \frac{\partial P_z}{\partial x} + g \int_{z_b}^z \frac{\partial \rho}{\partial x} dz - g \rho_b \frac{\partial z_b}{\partial x}.$$

When $z = z_b$ the second term on the right is eliminated so that

$$\frac{\partial P_b}{\partial x} = \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b} - g \rho_b \frac{\partial z_b}{\partial x}.$$

With this substitution the terms from (13a) become

$$-\frac{P_b \alpha_0(P_b)}{g} \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b} + P_b \alpha_0(P_b) \rho_b \frac{\partial z_b}{\partial x} - P_b \frac{\partial z_b}{\partial x}.$$

It is assumed that $\alpha_0(P_b) = \alpha_b$ and the bottom pressure and slope terms of (13) may be represented by the single term $-\frac{P_b \alpha_b}{g} \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b}$

To be useful for the present purposes it is necessary to neglect additional terms in (13a, b). Assume (a) that field acceleration terms are negligible; (b) that the velocities near the bottom are very small so that τ_b is much smaller than τ_s , so the vertical eddy viscosity terms when integrated over the column are closely approximated by τ_s ; and (c) that in a central region of the ocean horizontal friction terms are negligible (Munk, 1950). With

these assumptions, the transport equations may be written:

$$\begin{aligned} -fV &= -\frac{\partial \chi}{\partial x} - \frac{P_b \alpha_b}{g} \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b} + \tau_{sx}, \\ fU &= -\frac{\partial \chi}{\partial y} - \frac{P_b \alpha_b}{g} \left(\frac{\partial P_z}{\partial y} \right)_{z=z_b} + \tau_{sy}. \end{aligned} \quad (14a, b)$$

There are no existing methods of determining open ocean transports with sufficient precision to account for mass exchange at the sea surface by continuity methods. To a sufficient degree of accuracy the divergence of the transport is expressed by

$$\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} = 0. \quad (15)$$

Equations (14a, b) and (15) are the basic ones for this problem.

After Fofonoff (1962), the components U and V of the transport can be separated into barotropic, baroclinic, and Ekman modes:

$$U = U_b + U_g + U_E \quad (16a, b)$$

$$V = V_b + V_g + V_E$$

The barotropic mode is interpreted as a velocity, v_b , constant throughout the water column, so that

$$\begin{aligned} fu_b &= -\alpha_b \left(\frac{\partial P_z}{\partial y} \right)_{z=z_b}, & fv_b &= \alpha_b \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b}; \\ U_b &= \int_{z_b}^{\eta} \rho u_b dz, & V_b &= \int_{z_b}^{\eta} \rho v_b dz. \end{aligned} \quad (17a, b)$$

since u_b and v_b are not functions of z

$$U_b = u_b \int_{z_b}^{\eta} \rho dz, \quad V_b = v_b \int_{z_b}^{\eta} \rho dz.$$

By the hydrostatic equation

$$\begin{aligned} U_b &= \frac{u_b}{g} \int_{P_o}^{P_b} dP, & V_b &= \frac{v_b}{g} \int_{P_o}^{P_b} dP; \\ U_b &= \frac{u_b P_b}{g}, & V_b &= \frac{v_b P_b}{g} \end{aligned}$$

By substituting for u_b and v_b in equations (17a, b)

$$-fU_b = \frac{\alpha_b P_b}{g} \left(\frac{\partial P_z}{\partial y} \right)_{z=z_b}, \quad fV_b = \frac{\alpha_b P_b}{g} \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b} \quad (18a, b)$$

The baroclinic mode is interpreted as that part of the transport that is a consequence of pressure gradients arising from field variations in the density structure of the water. The velocities associated with this mode of transport are the relative velocities resulting from dynamic height computations. This mode of transport can be related to the pressure

terms on the right side of equations (14a, b) as follows:

$$\rho f(v_b + v_g) = \frac{\partial P}{\partial x},$$

and integrating

$$fV_b + fV_g = \int_{z_b}^{\eta} \frac{\partial P}{\partial x} dz.$$

Substituting for V_b from (18b),

$$\frac{\alpha_b P_b}{g} \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b} + fV_g = \int_{z_b}^{\eta} \frac{\partial P}{\partial x} dz.$$

Then using (11a) and the arguments following (13a, b) regarding bottom pressure terms,

$$\frac{\alpha_b P_b}{g} \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b} + fV_g = \frac{\partial \chi}{\partial x} + \frac{\alpha_b P_b}{g} \left(\frac{\partial P_z}{\partial x} \right)_{z=z_b}. \quad (19)$$

Identical pressure terms can be eliminated from the two sides of this expression. This, together with a similar development for U_g yields

$$fU_g = -\frac{\partial \chi}{\partial y}, \quad fV_g = \frac{\partial \chi}{\partial x}. \quad (20a, b)$$

The components of the Ekman transport are defined by the equations

$$fU_E = \tau_{sy}, \quad fV_E = -\tau_{sx}. \quad (21a, b)$$

Although the sum of the divergences of the three modes must be zero (15), the divergences of each mode taken separately need not be zero. Expressions for the individual divergences can be derived by cross differentiating the equations that define the barotropic, baroclinic, and Ekman transports—(18a, b), (20a, b), and (21a, b). Cross differentiating leads to

$$f \left(\frac{\partial U_b}{\partial x} + \frac{\partial V_b}{\partial y} \right) + \beta V_b = -\rho' f \left(u_b \frac{\partial z_b}{\partial x} + v_b \frac{\partial z_b}{\partial y} \right), \quad (22)$$

$$f \left(\frac{\partial U_g}{\partial x} + \frac{\partial V_g}{\partial y} \right) + \beta V_g = 0, \quad (23)$$

$$f \left(\frac{\partial U_E}{\partial x} + \frac{\partial V_E}{\partial y} \right) + \beta V_E = \frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y}. \quad (24)$$

In equation (22) ρ' is closely approximated by the mean density of the column.

The right side of equation (22) can be neglected when the gradient of the bottom is sufficiently small or when flow along the bottom is nearly normal to the slope of the bottom. (Applicability of this approximation is examined in section 2.2.) With this approximation, on adding equations (22), (23), and (24)

$$f \left[\frac{\partial (U_b + U_g + U_E)}{\partial x} + \frac{\partial (V_b + V_g + V_E)}{\partial y} \right] + \beta [V_b + V_g + V_E] = \frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y},$$

and on substituting for transports from equation (16a, b), and noting the continuity conditions expressed by (15)

$$\beta V = \frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y}. \quad (25)$$

This is the expression derived by Sverdrup (1947).

Equation (15) allows the introduction of a transport function, ψ , such that

$$U = \frac{\partial \psi}{\partial y}; \quad V = -\frac{\partial \psi}{\partial x}. \quad (26a, b)$$

Substituting from (26b) into (25) and integrating yields

$$\psi = -\frac{1}{\beta} \int \left(\frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y} \right) dx + C. \quad (27)$$

Using conditions at the eastern boundary, the constant

$$C = \frac{1}{\beta} \int_0^L \left(\frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y} \right) dx;$$

on combining with (27)

$$\psi = \frac{1}{\beta} \int_x^L \left(\frac{\partial \tau_{sy}}{\partial x} - \frac{\partial \tau_{sx}}{\partial y} \right) dx. \quad (28)$$

This solution (Fofonoff, 1962) does not provide for an eastern boundary layer. Fofonoff has justified the elimination of an eastern boundary layer by writing Munk's (1950) general solution for mass transport in a nondimensional form and simplifying the equation on the basis of relative orders of magnitude. For the area under consideration (the central North Pacific Ocean) provision for an eastern boundary layer is not necessary.

As mentioned above Sverdrup (1947), Munk (1950), and Fofonoff (1962) all specified that the transport was made up of baroclinic and Ekman modes only; the problem of treating convergence due to bathymetric effects on the barotropic flow thus was not pertinent.

Fofonoff (1960) used equation (28) (in geographic coordinates) as the basis for his mass transport computations in the Pacific. Using monthly means of sea-level atmospheric pressure (from the records of the U.S. Weather Bureau) he successively computed:

- geostrophic wind velocities,
- surface wind velocities,
- wind stress on the surface,
- meridional and zonal Ekman transport, V_E and U_E ,
- meridional total transport, V ,
- integrated total transport, ψ , and
- integrated geostrophic transport, ψ_g .

The pressure distribution and each of the components or modes of transport are tabulated on alternate five-degree grid points for the Pacific Ocean north of latitude 20°N. Tabulations have been published for each month during the period January 1950 through May 1962 (Fofonoff 1960, 1961; Fofonoff and Ross, 1962; Fofonoff and Dobson, 1963). Similar tabulations have been made for months of interest during the period June 1962 through

June 1963 by the Applied Mathematics staff of the Department of Oceanography, University of Washington.

Transformed into geographic coordinates the quantities of interest are:

$$V_E = -\frac{\tau_\lambda}{f}, \quad U_E = \frac{\tau_\phi}{f}, \quad (29a, b)$$

$$V = \left(\frac{\partial \tau_\phi}{\partial \lambda} - \frac{\partial(\tau_\lambda \cos \phi)}{\partial \phi} \right) \frac{1}{\beta R \cos \phi} \quad (30)$$

The transport function, ψ , is defined by

$$U = \frac{1}{R} \frac{\partial \psi}{\partial \phi}, \quad V = -\frac{1}{R \cos \phi} \frac{\partial \psi}{\partial \lambda}. \quad (31a, b)$$

Equation (31b) is integrated from the eastern shoreline,

$$\psi = -\frac{1}{\beta} \int_{\lambda_0}^{\lambda} \left(\frac{\partial \tau_\phi}{\partial \lambda} - \frac{\partial(\tau_\lambda \cos \phi)}{\partial \phi} \right) d\lambda. \quad (32)$$

Fofonoff (1960) tabulates the geostrophic transport as a transport function, ψ_g . He notes this procedure is not entirely correct, since the geostrophic transport is not divergenceless. He obtains his values for ψ_g from the expression

$$\psi_g = \int_{\lambda_0}^{\lambda} \left[-\frac{1}{\beta} \left(\frac{\partial \tau_\phi}{\partial \lambda} - \frac{\partial(\tau_\lambda \cos \phi)}{\partial \phi} \right) + \frac{\tau_\lambda}{f} R \cos \phi \right] d\lambda, \quad (33)$$

that is, the difference $V - V_E$ integrated over longitude. Following this procedure the meridional geostrophic transport is accurately represented by ψ_g ; however, the zonal geostrophic transport is incorrectly represented since it reflects the effects of neglecting divergence of the Ekman transport.

Both the transport function for total transport and that for geostrophic transport can depict zonal transport in error as a consequence of using the *grid point* nearest the eastern boundary as the origin for integration. However, in the present use of Fofonoff's tabulated transports as reference values, results are insensitive to the wind-driven transport to such a degree that errors caused by unaccounted-for zonal transport near the eastern boundary will not greatly influence the final values.

With the values tabulated by Fofonoff (1960), baroclinic transport computed from oceanographic station data, and equation (16a, b), it is possible to obtain barotropic transport through a given section:

$$U_b = U - U_E - U_g, \quad (16a, b)$$

$$V_b = V - V_E - V_g.$$

In the present problem the intervals of interest are determined by the locations of the oceanographic stations. The deep stations used are located so as to form either zonal or meridional sections. In equations (16a, b) the first terms on the right can be extracted from Fofonoff's (1960) tabulated transport functions for total transport by finite differences; the second terms, Ekman transports, can be integrated over the interval from the tabulated values; and the third term can be taken from the oceanographic data. Equations (17a, b) can be closely approximated by

$$U_b \approx \rho_m u_b h, \quad (34)$$

$$V_b \approx \rho_m v_b h.$$

And so it is possible to solve for the barotropic velocities u_b, v_b

$$u_b = \frac{(\psi_{\phi_2} - \psi_{\phi_1}) - \sum_{\phi_1}^{\phi_2} U_E R \Delta\phi - \left(\frac{\chi_{\phi_2} - \chi_{\phi_1}}{f} \right)}{\rho_m h R \Delta\phi}$$

$$v_b = \frac{(\psi_{\lambda_2} - \psi_{\lambda_1}) - \sum_{\lambda_1}^{\lambda_2} V_E R \cos \phi \Delta\lambda - \left(\frac{\chi_{\lambda_2} - \chi_{\lambda_1}}{f} \right)}{\rho_m h R \cos \phi \Delta\lambda} \quad (35a, b)$$

These barotropic velocities, constant throughout the column, can be combined with baroclinic velocities based on observations over the entire depth of the column to obtain absolute geostrophic velocities.

Appendix II.

A Numerical Example

Based on equations (35a, b), velocity profiles are determined for the meridional and zonal sections defined by oceanographic stations 29, 32, 59, and 63, September 1961.

The first step in the determination is to extract the geostrophic mode of transport for the section of interest. This transport is included in equations (35a, b) as the first and last terms in the numerator on the right side. Values of the total transport function along the meridians of interest were plotted against latitude (fig. 64). The stations are indicated on the plots of ψ , and the total transport through the sections between the indicated stations is taken by difference:

Section

$$63-59 \quad (\psi_{\phi_b} - \psi_{\phi_a}) = [-137 - (-112)] \cdot 10^5 = -250 \cdot 10^4 \text{ metric tons/sec}$$

$$29-32 \quad (\psi_{\phi_b} - \psi_{\phi_a}) = [-192 - (-113)] \cdot 10^5 = -790 \cdot 10^4 \text{ metric tons/sec}$$

$$63-29 \quad (\psi_{\lambda_b} - \psi_{\lambda_a}) = [-113 - (-112)] \cdot 10^5 = -10 \cdot 10^4 \text{ metric tons/sec}$$

$$59-32 \quad (\psi_{\lambda_b} - \psi_{\lambda_a}) = [-192 - (-137)] \cdot 10^5 = -550 \cdot 10^4 \text{ metric tons/sec}$$

It can be seen that continuity requirements for the total flow are satisfied by the results of this procedure. Plots of zonal and meridional Ekman transport along meridians and parallels are shown as figures 65, 66, and 67. Again the stations are indicated on the plots; summations to obtain Ekman transport through the sections are performed using Simpson's rule, or the trapezoid rule:

Section

$$63-59 \quad \Sigma U_E R d\phi = \left[\frac{98(2+12+4)}{3} + \frac{142(4+36+15)}{3} \right] \cdot 10 \\ = 3.2 \cdot 10^4 \text{ metric tons/sec}$$

$$29-32 \quad \Sigma U_E R d\phi = \left[\frac{190(8.5+44+9.5)}{3} + \frac{52(9.5+34+8)}{3} \right] \cdot 10 \\ = 4.8 \cdot 10^4 \text{ metric tons/sec}$$

$$63-29 \quad \Sigma V_E R \cos \phi d\lambda = \left[\frac{344(18+94+36.5)}{3} + \frac{463(36.5+210+58.5)}{3} \right] \cdot 10 \\ = 64.1 \cdot 10^4 \text{ metric tons/sec}$$

$$59-32 \quad \Sigma V_E R \cos \phi d\lambda = \left[\frac{1520(2+(-2.2))}{2} \right] \cdot 10 \\ = -0.2 \cdot 10^4 \text{ metric tons/sec}$$

The wind-driven geostrophic transport for a given section is the difference between the total and Ekman transports:

Section

$$63-59 = (-250 - 3.2) \cdot 10^4 \approx -253 \cdot 10^4 \text{ metric tons/sec}$$

$$29-32 = (-790 - 4.8) \cdot 10^4 \approx -795 \cdot 10^4 \text{ metric tons/sec}$$

$$63-29 = (-10 - 64.1) \cdot 10^4 \approx -74 \cdot 10^4 \text{ metric tons/sec}$$

$$59-32 = (-550 - (-0.2)) \cdot 10^4 \approx -550 \cdot 10^4 \text{ metric tons/sec}$$

Because the distribution of horizontal velocities is not highly sensitive to this geostrophic transport, values for it can be extracted directly from tabulations of the geostrophic

transport stream function, ψ_g . (As noted in appendix I, this procedure eliminates from consideration the Ekman divergence.) Plots of the geostrophic transport function similar to those for the total transport function are made (fig. 68) and the geostrophic transport is extracted by difference. A comparison of alternate values of geostrophic transport for the four sections considered is given below.

<i>Section</i>	<i>Transport considering Ekman divergence</i>	<i>Transport from geostrophic transport function</i>
63-59	$-253 \cdot 10^4$ metric tons/sec	$-20 \cdot 10^4$ metric tons/sec
29-32	$-795 \cdot 10^4$ metric tons/sec	$-310 \cdot 10^4$ metric tons/sec
63-29	$-74 \cdot 10^4$ metric tons/sec	$-250 \cdot 10^4$ metric tons/sec
59-32	$-550 \cdot 10^4$ metric tons/sec	$-540 \cdot 10^4$ metric tons/sec

The dynamic topographies arising from a comparison of transport from dynamic heights adjusted to each of these sets of wind-driven transport are shown on figures 23 through 29. For the balance of the numerical example the transport including the effects of divergence is used.

The final terms in equations (35a, b) express mass transport through the sections due to internal density structure. To obtain this baroclinic transport, data from each station is submitted to a machine interpolation program which provides the following values at standard depths:

temperature,
salinity,
sigma-t,
specific volume anomaly, δ ,
geopotential anomaly, ΔD , referred to the surface,
anomaly of potential energy, χ ,
and values for other variables represented in the field data.

For convenience in the computations advantage is taken of the near numerical equality between mass transport and volume transports, T and T_E . The mass transports determined from the curl of the wind stress are entered in subsequent comparisons as though they were volume transports. The final terms in (35a, b) are determined from differences in geopotential anomalies. Equations (35a, b), reflecting these near-equalities, become:

$$u_b = \frac{T - T_E - \left(\frac{\Delta D_b - \Delta D_a}{f} \right)}{hR\Delta\phi},$$

$$v_b = \frac{T - T_E - \left(\frac{\Delta D_b - \Delta D_a}{f} \right)}{hR \cos \phi \Delta\lambda}. \quad (36a, b)$$

Next, values are determined for the length-of-section terms, $R \cos \phi \Delta\lambda$ and $R\Delta\phi$, and for $\frac{1}{f}$, the reciprocal of the Coriolis parameter:

<i>Section</i>	<i>Mid-latitude</i>	<i>Length (km)</i>	$\frac{1}{f} (10^5 \text{ sec})$
63-59	34°50'	480	0.1201
29-32	34°33'	485	0.1209
63-29	32°31'	1613	0.1276
59-32	36°52'	1520	0.1141

The depth of the column, h , was approximated as 5,000 meters, and geopotential anomalies from the station data were referred to the same depth. Observations extended to at least 5,000 meters at most of the stations used in the program; at stations where they did not, values of temperature and salinity were selected to extend the profiles to 5,000 meters. These values were selected by extrapolating profiles at the station considered and by interpolating between appropriate values at adjacent stations. In the selection care was taken to avoid creating artificial structure in the water mass.

Values of the wind-driven transport, geopotential anomalies referred to 5,000 meters, the distance between stations, and the reciprocal of the Coriolis parameter are submitted to a second program. This program calculates velocities and geopotential anomalies at each standard depth. These velocities and anomalies reflect the variation with depth associated with the structure of the water column and are adjusted to describe a transport through the section equal to the geostrophic mode of transport from the curl of wind stress—that is, equations (36a, b) are satisfied. Tables 2 through 5 show the data printouts from this program for the four sections considered. An explanation of the quantities shown in the headings and columns of each table follows.

Quantities in the headings are:

STATIONS, identifying numbers of oceanographic stations

L, length of section in kilometers

COR, reciprocal of the Coriolis parameter, multiplied by 10^{-5}

TRANSPORT, the wind-driven geostrophic transport in 10^4 m³/second

N, the number of standard depths considered

SUMVZ, transport in 10^4 m³/second as obtained from differences of dynamic heights, referred to 5,000 meters; the sum of the values in column VVZ below

CORR, a correction factor in volume transport per meter of depth in 10^2 m²/second, obtained by:

$$\text{CORR} = \frac{\text{TRANSPORT} - \text{SUMVZ}}{h}, \text{ that is}$$

$$\text{CORR} = \frac{T - T_E - \frac{\Delta D_b - \Delta D_a}{f}}{h}.$$

SUMVZ*, a check of computations. The value is identical with that for TRANSPORT.

Quantities in the columns are:

DEPTH, standard depths

A, geopotential anomalies in dynamic meters for station A, referred to 5,000 meters; station A is the southern station of meridional sections and the eastern station of the pairs along parallels of latitude.

B, geopotential anomalies for station B

C, the difference, column A minus column B at each standard depth

VZ, volume transport per meter of depth in 10^2 m²/second applicable to each depth.

This function of volume transport is obtained by multiplying the difference in geopotential anomalies (column C) by the reciprocal of the Coriolis parameter (COR in heading).

VVZ, baroclinic transport in 10^4 m³/second for each depth interval, obtained by numerical integration of VZ with respect to depth

VCZ, an adjusted volume per second per meter of depth applicable to each depth.

VCZ is obtained by adding VZ at each depth to the correction, CORR, as given in the heading of the table.

V*, the velocity at each standard depth corresponding to the adjusted transport values,

VCZ. These velocities, obtained in cm/second, are the absolute geostrophic

velocities. The value of V^* at the 5,000-meter level corresponds to u_b or v_b in equations (36a, b).

C^* , adjusted differences in geopotential anomaly for the section. These adjusted differences are accepted as reflections of the absolute geopotential topography.

TABLE 2. — *Data printout for section defined by oceanographic stations 63 and 59*

STATIONS 063-059									
L=480. KMS		COR=0.1201		TRANSPORT=-253.		E4 M3/SEC		E2 M2/SEC	
N=24 DEPTHS		SUMVZ=160.8438		E4 M3/SEC		CORR=-8.2769			
SUMVZ*=-253.0000									
DEPTH	A	B	C	VZ	VVZ	VCZ	V*	C*	
0	3.7660	3.7660	0.	0.	0.1201	-8.2769	-0.1724	-0.0069	
10	3.7230	3.7210	0.0020	2.4020	0.3603	-5.8749	-0.1224	-0.0049	
20	3.6810	3.6770	0.0040	4.8040	0.6606	-3.4729	-0.0724	-0.0029	
30	3.6390	3.6320	0.0070	8.4070	3.1226	0.1301	0.0027	0.0001	
50	3.5710	3.5520	0.0190	22.8190	6.9058	14.5421	0.3030	0.0121	
75	3.5050	3.4780	0.0270	32.4270	7.6564	24.1501	0.5031	0.0201	
100	3.4460	3.4220	0.0240	28.8240	11.7098	20.5471	0.4281	0.0171	
150	3.3340	3.3190	0.0150	18.0150	6.3052	9.7381	0.2029	0.0081	
200	3.2260	3.2200	0.0060	7.2060	1.5012	-1.0709	-0.0223	-0.0009	
250	3.1230	3.1240	-0.0010	-1.2010	-2.4020	-9.4779	-0.1975	-0.0079	
300	3.0240	3.0310	-0.0070	-8.4070	-12.0100	-16.6839	-0.3476	-0.0139	
400	2.8410	2.8540	-0.0130	-15.6130	-16.2135	-23.8899	-0.4977	-0.0199	
500	2.6770	2.6910	-0.0140	-16.8140	-15.6130	-25.0909	-0.5227	-0.0209	
600	2.5330	2.5450	-0.0120	-14.4120	-12.0100	-22.6889	-0.4727	-0.0189	
700	2.4070	2.4150	-0.0080	-9.6080	-7.2060	-17.8849	-0.3726	-0.0149	
800	2.2960	2.3000	-0.0040	-4.8040	-3.6030	-13.0809	-0.2725	-0.0109	
1000	2.1060	2.1050	0.0010	1.2010	6.0050	-7.0759	-0.1474	-0.0059	
1200	1.9450	1.9410	0.0040	4.8040	14.4120	-3.4729	-0.0724	-0.0029	
1500	1.7330	1.7290	0.0040	4.8040	27.0225	-3.4729	-0.0724	-0.0029	
2000	1.4350	1.4300	0.0050	6.0050	42.0349	-2.2719	-0.0473	-0.0019	
2500	1.1840	1.1750	0.0090	10.8090	48.0400	2.5321	0.0528	0.0021	
3000	0.9490	0.9420	0.0070	8.4070	48.0400	0.1301	0.0027	0.0001	
4000	0.4850	0.4840	0.0010	1.2010	6.0050	-7.0759	-0.1474	-0.0059	
5000	0.	0.	0.	0.	0.	-8.2769	-0.1724	-0.0069	

TABLE 3. — Data printout for section defined by oceanographic stations 29 and 32

DEPTH	STATIONS 029—032									
	L = 485. KMS		COR = 0.1209		TRANSPORT = -795.		E4 M3/SEC		E4 M3/SEC	
	N = 24 DEPTHS	SUMVZ = -141.7854	C	VZ	VVZ	CORR = -13.0643	E2 M2/SEC	V*	C*	
					SUMVZ* = -795.0000					
	A	B	C	VZ	VVZ	VCZ				
0	3.7310	3.5890	0.1420	171.6780	16.6842	158.6137		3.2704	0.1312	
10	3.6820	3.5480	0.1340	162.0060	15.7170	148.9417		3.0710	0.1232	
20	3.6330	3.5070	0.1260	152.3340	14.7498	139.2697		2.8715	0.1152	
30	3.5840	3.4660	0.1180	142.6620	27.9279	129.5977		2.6721	0.1072	
50	3.5080	3.3950	0.1130	136.6170	33.7009	123.5527		2.5475	0.1022	
75	3.4410	3.3310	0.1100	132.9900	32.0385	119.9257		2.4727	0.0992	
100	3.3810	3.2790	0.1020	123.3180	56.8230	110.2537		2.2733	0.0912	
150	3.2670	3.1810	0.0860	103.9740	47.4532	90.9097		1.8744	0.0752	
200	3.1580	3.0870	0.0710	85.8390	38.6880	72.7747		1.5005	0.0602	
250	3.0540	2.9970	0.0570	68.9130	30.8295	55.8487		1.1515	0.0462	
300	2.9550	2.9100	0.0450	54.4050	43.5240	41.3407		0.8524	0.0342	
400	2.7720	2.7450	0.0270	32.6430	24.7845	19.5787		0.4037	0.0162	
500	2.6120	2.5980	0.0140	16.9260	10.8810	3.8617		0.0796	0.0032	
600	2.4730	2.4690	0.0040	4.8360	-1.2090	-8.2283		-0.1697	-0.0068	
700	2.3490	2.3550	-0.0060	-7.2540	-13.9035	-20.3183		-0.4189	-0.0168	
800	2.2370	2.2540	-0.0170	-20.5530	-54.4050	-33.6173		-0.6931	-0.0278	
1000	2.0470	2.0750	-0.0280	-33.8520	-71.3310	-46.9163		-0.9673	-0.0388	
1200	1.8900	1.9210	-0.0310	-37.4790	-110.6235	-50.5433		-1.0421	-0.0418	
1500	1.6900	1.7200	-0.0300	-36.2700	-151.1250	-49.3343		-1.0172	-0.0408	
2000	1.4100	1.4300	-0.0200	-24.1800	-87.6525	-37.2443		-0.7679	-0.0308	
2500	1.1660	1.1750	-0.0090	-10.8810	-33.2475	-23.9453		-0.4937	-0.0198	
3000	0.9370	0.9390	-0.0020	-2.4180	-12.0900	-15.4823		-0.3192	-0.0128	
4000	0.4770	0.4770	0.	0.	0.	-13.0643		-0.2694	-0.0108	
5000	0.	0.	0.	0.	0.	-13.0643		-0.2694	-0.0108	

TABLE 4.—Data printout for section defined by oceanographic stations 63 and 29

STATIONS 063—029

L = 1613. KMS COR = 0.1276 TRANSPORT = -74. E4 M3/SEC
N = 24 DEPTHS SUMVZ = 1761.1349 E4 M3/SEC CORR = -36.7027 E2 M2/SEC

SUMVZ* = -73.9999

DEPTH	A	B	C	VZ	VVZ	VCZ	V*	C*
0	3.7660	3.7310	0.0350	44.6600	4.8488	7.9573	0.0493	0.0062
10	3.7230	3.6820	0.0410	52.3160	5.6782	15.6133	0.0968	0.0122
20	3.6810	3.6330	0.0480	61.2480	6.5714	24.5453	0.1552	0.0192
30	3.6390	3.5840	0.0550	70.1800	15.0568	33.4773	0.2075	0.0262
50	3.5710	3.5080	0.0630	80.3880	20.2565	43.6853	0.2708	0.0342
75	3.5050	3.4410	0.0640	81.6640	20.5755	44.9613	0.2787	0.0352
100	3.4460	3.3810	0.0650	82.9400	42.1080	46.2373	0.2867	0.0362
150	3.3340	3.2670	0.0670	85.4920	43.0650	48.7893	0.3025	0.0382
200	3.2260	3.1580	0.0680	86.7680	43.7030	50.0653	0.3104	0.0392
250	3.1230	3.0540	0.0690	88.0440	44.0220	51.3413	0.3183	0.0402
300	3.0240	2.9550	0.0690	88.0440	88.0440	51.3413	0.3183	0.0402
400	2.8410	2.7720	0.0690	88.0440	85.4920	51.3413	0.3183	0.0402
500	2.6770	2.6120	0.0650	82.9400	79.7500	46.2373	0.2867	0.0362
600	2.5330	2.4730	0.0600	76.5600	75.2840	39.8573	0.2471	0.0312
700	2.4070	2.3490	0.0580	74.0080	74.6460	37.3053	0.2313	0.0292
800	2.2960	2.2370	0.0590	75.2840	150.5680	38.5813	0.2392	0.0302
1000	2.1060	2.0470	0.0590	75.2840	145.4640	38.5813	0.2392	0.0302
1200	1.9450	1.8900	0.0550	70.1900	187.5720	33.4773	0.2075	0.0262
1500	1.7330	1.6900	0.0430	54.8680	216.9200	18.1653	0.1126	0.0142
2000	1.4350	1.4100	0.0250	31.9000	137.1699	-4.8027	-0.0298	-0.0038
2500	1.1840	1.1660	0.0180	22.9680	95.7000	-13.7347	-0.0852	-0.0108
3000	0.9490	0.9370	0.0120	15.3120	127.6000	-21.3907	-0.1326	-0.0168
4000	0.4850	0.4770	0.0080	10.2080	51.0400	-26.4947	-0.1643	-0.0208
5000	0.	0.	0.	0.	0.	-36.7027	-0.2275	-0.0288

TABLE 5.—Data printout for section defined by oceanographic stations 59 and 32

STATIONS 059—032									
L= 1520. KMS		COR=0.1143		TRANSPORT= —550.		E4 M3/SEC		E2 M2/SEC	
N= 24 DEPTHS		SUMVZ= 1290.4469		E4 M3/SEC		CORR= —36.8089			
SUMVZ* = —549.9999									
DEPTH	A	B	C	VZ	VVZ	VCZ	V*	C*	
0	3.7660	3.5890	0.1770	202.3110	20.0025	165.5020	1.0888	0.1448	
10	3.7210	3.5480	0.1730	197.7390	19.6024	160.9301	1.0588	0.1408	
20	3.6770	3.5070	0.1700	194.3100	19.2024	157.5010	1.0362	0.1378	
30	3.6320	3.4660	0.1660	189.7380	36.9189	152.9291	1.0061	0.1338	
50	3.5520	3.3950	0.1570	179.4510	43.4340	142.6421	0.9384	0.1248	
75	3.4780	3.3310	0.1470	168.0210	41.4338	131.2121	0.8632	0.1148	
100	3.4220	3.2790	0.1430	163.4490	80.2957	126.6401	0.8332	0.1108	
150	3.3190	3.1810	0.1380	157.7340	77.4382	120.9250	0.7956	0.1058	
200	3.2200	3.0870	0.1330	152.0190	74.2950	115.2101	0.7580	0.1008	
250	3.1240	2.9970	0.1270	145.1610	70.8660	108.3521	0.7128	0.0948	
300	3.0310	2.9100	0.1210	138.3030	131.4450	101.4941	0.6677	0.0888	
400	2.8540	2.7450	0.1090	124.5870	115.4430	87.7781	0.5775	0.0768	
500	2.6910	2.5980	0.0930	106.2990	96.5835	69.4901	0.4572	0.0608	
600	2.5450	2.4690	0.0760	86.8680	77.7240	50.0591	0.3293	0.0438	
700	2.4150	2.3550	0.0600	68.5800	60.5790	31.7711	0.2090	0.0278	
800	2.3000	2.2540	0.0460	52.5780	86.8680	15.7691	0.1037	0.0138	
1000	2.1050	2.0750	0.0300	34.2900	57.1500	—2.5189	—0.0166	—0.0022	
1200	1.9410	1.9210	0.0200	22.8600	49.7205	—13.9489	—0.0918	—0.0122	
1500	1.7290	1.7200	0.0090	10.2870	25.7175	—26.5220	—0.1745	—0.0232	
2000	1.4300	1.4300	0.	0.	0.	—36.8089	—0.2422	—0.0322	
2500	1.1750	1.1750	0.	0.	8.5725	—36.8089	—0.2422	—0.0322	
3000	0.9420	0.9390	0.0030	3.4290	57.1500	—33.3799	—0.2196	—0.0292	
4000	0.4840	0.4770	0.0070	8.0010	40.0050	—28.8079	—0.1895	—0.0252	
5000	0.	0.	0.	0.	0.	—36.8089	—0.2422	—0.0322	

Appendix III.

List of Symbols

A	coefficient of eddy viscosity	α	specific volume
C_D	drag coefficient	α_o	specific volume of sea water, salinity 35 $^{\circ}$ / $_{\infty}$, temperature $^{\circ}$ C, and pressure p .
E_p	potential energy anomaly	β	change of Coriolis parameter with latitude
f	Coriolis parameter	δ	specific volume anomaly
g	acceleration of gravity	η	elevation of the sea surface
h	depth of water	λ	angle of longitude
L	length of section	ρ	density
P	hydrostatic pressure	τ	shearing stress
R	Earth's radius	ϕ	angle of latitude
T	volume transport	χ	anomaly of potential energy
u, v, w	components of velocity	ψ	mass transport function
x, y, z	cartesian coordinates	∇^2	Laplacian operator
ΔD	geopotential anomaly	ω	Earth's angular velocity
V	mass transport per unit width		
θ	angle streamlines make with bottom slope		

Subscripts:

b	barotropic, bottom
E	Ekman
g	baroclinic

FIGURE 1. — Locations of stations and sections.

FIGURE 2.—Stations used in determination of transport.

LONGITUDE 177° W

FIGURE 3.—Temperature (°C) Section 1, September 1961.

FIGURE 4. — Salinity (‰) Section 1, September 1961.

FIGURE 5.—Dissolved oxygen (ml/L) Section 1, September 1961.

FIGURE 6.—Temperature (°C) Section 2, September 1961.

FIGURE 8.—Dissolved oxygen (ml/L) Section 2, September 1961.

FIGURE 11. — Dissolved oxygen (ml/L) Section 3, September 1962.

FIGURE 12.—Temperature (°C) Section 4, September 1963.

FIGURE 13.—Salinity (‰) Section 4, September 1963.

FIGURE 14.—Dissolved oxygen (ml/L) Section 4, September 1963.

FIGURE 15.—Temperature (°C) Section 5, May 1963.

FIGURE 16.—Salinity (‰) Section 5, May 1963.

FIGURE 17.—Dissolved oxygen (ml/L) Section 5, May 1963.

FIGURE 18. — Variation with latitude of wind-driven integrated geostrophic transport, September 1962.

FIGURE 19.—Geopotential anomaly at 4,000 meters referred to 5,000 meters, Sections 2, 3, and 5.

FIGURE 20. — Geopotential anomaly at 4,000 meters referred to 5,000 meters, Sections 1 and 4.

FIGURE 21. — Geopotential anomaly at 3,000 meters referred to 5,000 meters, Sections 2, 3, and 5.

FIGURE 22.—Geopotential anomaly at 3,000 meters referred to 5,000 meters, Sections 1 and 4.

FIGURE 23.—Topography in dynamic meters, 250-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 24.—Topography in dynamic meters, 500-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 25.—Topography in dynamic meters, 1,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 26.—Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 27. — Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 28.—Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 29. — Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, September 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 30.—Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, August 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 31.—Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, August 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 32.—Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, August 1961. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 33.—Topography in dynamic meters, 250-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 34.—Topography in dynamic meters, 500-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 35.—Topography in dynamic meters, 1,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 36.—Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 37.—Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 38.—Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, September 1962. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 39.—Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, September 1962. Drag coefficient 0.0026; meaned stresses from twice-daily reports of pressure.

FIGURE 40.—Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, September 1962. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.

FIGURE 41. — Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, September 1962. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.

FIGURE 42.—Topography in dynamic meters, 250-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 43.—Topography in dynamic meters, 500-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 44.—Topography in dynamic meters, 1,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 45.—Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 46.—Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 47.—Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 48.—Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence not considered, May 1963. Drag coefficient 0.0026; stresses from mean monthly pressures.

FIGURE 49.—Topography in dynamic meters, 250-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.

FIGURE 50.—Topography in dynamic meters, 500-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.

FIGURE 51.—Topography in dynamic meters, 1,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.

FIGURE 52.—Topography in dynamic meters, 2,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.

FIGURE 53.—Topography in dynamic meters, 4,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.

FIGURE 54.—Topography in dynamic meters, 5,000-m level adjusted to wind-driven geostrophic transport, Ekman divergence considered, May 1963. Drag coefficient 0.0012; mean stresses from twice-daily reports of pressure.

FIGURE 55. — Variation of vertical velocity with depth at indicated latitudes along approximately 170° W.

FIGURE 56.—Depth of no zonal motion, 160° W, 1961.

FIGURE 57.—Depth of no zonal motion, 177° W, 1961.

FIGURE 58.—Depth of no meridional motion, 169° W, 1961.

FIGURE 59. — Depth of no zonal motion, 160° W, 1962.

Latitude

FIGURE 60.—Depth of no zonal motion, 165° W, May 1963.

FIGURE 61. — Depth of no zonal motion, 180° W, May 1963.

FIGURE 62. — Depth of no meridional motion, 172° W, May 1963.

FIGURE 63.— Annual variations in zonal velocity at 4,000 m.

FIGURE 64.—Variation with latitude of integrated total transport, September 1961.

FIGURE 65. —Variation with latitude of zonal Ekman transport, September 1961.

FIGURE 66.—Variation with latitude of zonal Ekman transport, September 1961.

FIGURE 67. — Variation with longitude of meridional Ekman transport, September 1961.

Latitude

FIGURE 68. — Variation with latitude of integrated geostrophic transport, September 1961.

MARINE
BIOLOGICAL
LABORATORY

LIBRARY

WOODS HOLE, MASS.
W. H. O. I.