SOME OCEANOGRAPHIC FEATURES OF THE NORTHEAST
PACIFIC OCEAN DURING AUGUST 1955

by

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Abstract

Physical oceanographic data from the international NorPac survey of August 1955, in the area north of Latitude 45° N and east of Longitude 161° W, were examined. The temperature, salinity, and density distribution and structure from the surface to 2000 meters are discussed.

The temperature structure showed an isothermal layer to about 30 meters depth, a marked thermocline to about 100 meters depth, a temperature inversion in most of the area, and below this a gradual temperature decrease into the abyss. At all depths the water was coldest in a "cold core" centered about 100 miles south of Kodiak and the Shumagin Islands. From there the temperature increased at each level in all directions.

The salinity structure showed an isohaline layer to about 100 meters depth, a marked halocline to about 200 meters depth, and below this the salinity increased slightly into the abyss. The salinity structure did not coincide with the temperature structure.

The density structure showed an isopycnal layer to about 30 meters depth, a pycnocline associated with the thermocline, a second isopycnal layer, a second pycnocline associated with the halocline, and below this the density increased slightly into the abyss.

Variations in these structures throughout the region are discussed in some detail.

There is no horizontal isosteric level in the 2000 meters of depth. It is concluded that there is no level of "no net motion" in this range, but a reference level of 2000 decibars for dynamic calculations is more acceptable than the usual 1000 decibar level. A new procedure

is introduced to extend the reference level into the bottom in near coastal areas.

The geostrophic currents were calculated. There was a major latitudinal drift from the west into the central part of the area. It veered northward and continued around the Gulf of Alaska, forming the Alaska Gyral, and left the area to the westward, as an intensified current (Alaska Stream) close along the Alaskan Peninsula. This intensification is probably due to conservation of absolute vorticity through changing latitude.

The circulation pattern extended to at least 2000 meters depth, and probably to the bottom. It transported about 17 million cubic meters of water per second. There were a number of eddies in the system, some of which were observed on earlier surveys.

The major flow pattern was not wind-generated within the region. The influence of local winds was limited to the upper 200 meters of depth. In some areas it aided the flow, and in others retarded it.

There is evidence to show that two chains of sea mounts influenced the current pattern according to the Bjerknes concept. Since the major portion of these is below 2000 meters depth it is concluded that currents exist in the abyss, in essentially the same direction as at the upper levels. This is consistent with the concept of conservation of absolute vorticity with changing latitude.

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INTRODUCTION

From late July to early September of 1955 the international NorPac oceanographic survey gathered data from the Pacific Ocean, north of Latitude 20° N. Since that time the Pacific Oceanographic Group of the Fisheries Research Board of Canada has undertaken several expeditions in further study of that region north of Latitude 40° N and east of Longitude 175° W. The intended result of these surveys is a complete description, both in time and space, of the oceanography of the Gulf of Alaska and adjacent waters. This report constitutes an integral part of that description and follows the initial preliminary assessment by Tully and Dodimead (1957).

This thesis concerns a description of the water structure and an interpretation of the magnitude of the volume transport of water interpreted from the 1955 NorPac data.

The area of investigation (hereafter referred to as the "region") (Figure 1) is that part of the Northeast Pacific Ocean north of Latitude 45° N and east of Longitude 161° W. The oceanographic stations included are:

- (a) University of Washington Stations 110-1 to 110-69. For facility, the cruise number will be dropped and these stations will be referred to as U of W Station 1, etc.
- (b) Pacific Oceanographic Group Stations 1 to 43 and 66 to 85 (POG Station 1, etc.)
- (c) Scripps Institute of Oceanography Stations 1 to 11 (SIO Station 1, etc.)

Figure 2a shows the location of these stations.

For the preliminary discussion of water structure, all one hundred and forty-two stations were used where possible, but in the section on volume transport only the forty-eight stations were used where the depth of observation reached, or could be extrapolated to, 2000 meters.

CHAPTER I

THE STRUCTURE AND PHYSICAL PROPERTIES OF THE WATER

The calculated geostrophic volume transport of water depends on the pressure (density) field. In turn, the density field is computed from the observed fields of temperature and salinity. In that which follows, the essential features of the temperature and salinity fields which existed in the region during August, 1955, are discussed, after which the corresponding density field is reviewed.

Three aspects of each field are investigated. First, the characteristics of the distribution with depth of each property is discussed at a representative station and then for the whole region. Next, the distribution of that property in six vertical sections is considered. Lastly, the lateral distribution of the property at several depths is reviewed.

The six traverses of the region and the included stations chosen for the vertical sections are shown in Figure 2b. The sections are centered approximately on Latitudes $54^{\circ} 20' N$, $51^{\circ} 00' N$ and $47^{\circ} 00' N$, and on Longitudes $161^{\circ} 00' W$, $147^{\circ} W$, and $135^{\circ} W$. For convenience, these sections are named Section 54, Section 51, Section 47, Section 161, Section 147, and Section 135, respectively. For any section the drawing of isopleths of any property was accomplished by reading, from the property versus depth curve (Station plot) for each station of the section, the depths at which the isopleths occurred. Then the depths common to one isopleth were joined by a smooth curve. For temperature sections,

bathythermograms, in the role of supplementary data, were used to place isotherms between stations, as well as to define the vertical temperature structure between serial observations at one station.

The surfaces selected for the discussion of the lateral distribution of a property are 10, 50, 100, 200, 400, 600 and 1000 meters depth. Values for each property at any depth were derived from the corresponding station plots.

Both the vertical sections and the lateral distributions are dependent on the station plots for values of any property. Therefore the three representations of data are concurrent, and as such any "smoothing" of one representation had a corresponding effect in the other two representations, but the serial observations of the parent station plot were never violated.

THE TEMPERATURE FIELD

Characteristics of the vertical distribution of temperature. A temperature versus depth curve, typical of these data from the northeastern Pacific Ocean, is shown in Figure 3. From the surface down, the curve exhibits in order:

- A. a shallow isothermal layer;
- B. a shallow layer in which temperature decreases rapidly (thermocline);
- C-D. a layer in which a temperature inversion may occur;
a lower zone in which temperature decreases slowly to the ocean floor.

Each layer is further discussed below.

The surface isothermal layer is the result of mixing by the surface wind. Figure 4 shows the depth of this isothermal layer over the region. The observed range of depth was 0 meters to 52 meters and the average of observed values was about 23 meters. Northward of Latitude 51° N the layer depth generally decreased from south to north, and from west to east, though a tongue with depths less than 20 meters extended westward from the Queen Charlotte Islands halfway across the region. An area of maximum layer depth (depths greater than 40 meters) lay on Latitude 51° N and extended from Longitude 154° W to Longitude 143° W. South and east of this was an area which had a surface layer depth of less than 10 meters. Southwest of the maximum area the depth decreased and was 13 meters at the corner of the region. Two areas at the coast had the surface layer less than 10 meters deep. One was located west of Vancouver

Island, the other, south of Kenai Peninsula.

The bottom of the surface isothermal layer (Layer Depth) marks the top of the thermocline. From there temperature decreased rapidly with depth to either a "minimum" value or to an "inflection point" value (these are discussed in the next paragraph). The temperature decrease in the thermocline is defined as the difference between the temperature at the bottom of the surface isothermal layer and the temperature of either the minimum point or inflection point of the temperature versus depth curves for each station. Illustrated in Figure 5 is the temperature decrease in the thermocline over the region. For the center portion of the region the decrease was about 6.5°C . South and east temperature decreases were larger, maxima of 9.5°C occurring on the southwest corner, and of 9.3°C about 50 miles southwest of Cape Flattery. North of the center portion the temperature decrease first became larger, reaching a maximum of 8.3°C about 200 miles southeast of Kodiak Island, and then became smaller to the coast. A minimum decrease of 4.6°C occurred about 50 miles southwest of the Shumagin Islands. A tongue of water of temperature decrease less than 6.0°C extended along Latitude 55°N from the coast to about Longitude 145°W . The average of all observed values of the decrease in the thermocline was 6.9°C .

The bottom of the thermocline was usually marked by a temperature minimum, i.e., in most cases there was a temperature increase below the thermocline. The origin of this temperature minimum has been explained by Tully (1953), from which the following is quoted¹:

¹ John P. Tully, "Some Characteristics of Sea Water Structure" (Nanaimo, B.C. Pacific Oceanographic Group, 1953), p. 21. (Multilithed).

"In March, towards the end of winter the upper zone waters are isothermal and at a minimum temperature, somewhat colder than the top of the deep zone. As the season advances the surface waters are warmed by vernal heating. This process may be regarded as the accumulation of heat from successive afternoon effects and wind mixing The warm upper zone continues to increase through Summer to mid-September. However, in the Autumn the surface cools and violent winter storms mix the waters. In this process the upper zone cools but continues to become deeper. Eventually its boundary intercepts the remnants of the previous winter boundary, and finally the characteristic late winter structure is formed."

This structure was named the dicothermal structure by Uda (1935).

Of the 111 stations considered which were deep enough to exhibit this feature, 102 (92 per cent) had the dicothermal structure. Therefore this is a characteristic feature of the vertical temperature structure of the water of the region. The temperature-depth curves of the other 9 stations had an inflection point at approximately the same depth at which the dicothermal structure usually appeared. The topography of the relative temperature minimum is illustrated in Figure 6a. By definition, this is also the topography of the lower limiting depth of the thermocline. The cross-hatched areas of Figure 6a indicate where the dicothermal structure was absent. Generally the depth of this feature was least near the coast (less than 100 meters). This was not true south of Cape St. Elias where the feature was found at depths exceeding 150 meters. The minimum depth at which it was observed is 60 meters

and that was 130 miles east of Kodiak Island. North of Latitude 52° N the depth of the temperature minimum was less than 125 meters with the marked exception of the northeast portion of the region, where at U of W Station 7 it occurred at 160 meters. The temperature minimum surface generally exceeded 125 meters in depth over the southern half of the region. A bi-tongue area showing a depth of greater than 150 meters lay in the south central part of the region.

Figure 6b shows the isotherm pattern on the dicothermal temperature surface. The minimum temperatures were less than 3.5° C and occurred in a small tongue about 100 miles south of the Shumagin Islands. From this "core" toward the coast or toward the southern limit of the region increasingly higher dico-temperatures were found. The isotherms 3.5° C, 4.0° C, 4.5° C and 5.0° C had the core pattern. The 5.5° C isotherm followed the 5.0° C isotherm on the south and east but turned toward the coast at Latitude 55° N. West of the Queen Charlotte Islands the 6.0° C isotherm appeared as a tongue. At the southeast corner of the region the dicothermal temperature was greater than 7.0° C. The maximum observed was 7.12° C.

Below the dicothermal layer a relative maximum temperature exists. Shown in Figure 7a is the topography of this maximum temperature. The observed range in depth was from 95 meters to 215 meters. The surface was shallowest near the coast except south of Cape St. Elias where it was as deep as 195 meters. Extending southwest from the Queen Charlottes almost to the southern limit of the region was a tongue of depth less than 150 meters. South of this tongue the depth of the maximum temperature increased to greater than 200 meters. West of the tongue

is an area of depth greater than 175 meters. A similar but smaller area occurred at Latitude 56° N, centered on U of W Station 7. In contrast, at the neighboring U of W Station 6 the maximum temperature was at 110 meters depth.

The isotherms on the maximum temperature surface are illustrated in Figure 7b. The core pattern was expressed by the 4.5° C and 5.0° C isotherms and was well supported by those of 5.5° C and 6.0° C, though the latter turned toward the coast at Latitude 54° N. Toward the southeast corner of the region the maximum temperature increased, reaching the observed maximum of 7.35° C about 80 miles southwest of Cape Flattery. Two areas of relative maximum temperature occurred, one west of the Queen Charlottes at POG Station 13 and the other northwest of that at U of W Station 6. This had a maximum temperature of 6.03° C while that of the neighboring U of W Station 7 was 4.06° C.

The difference between the relative maximum and minimum temperatures over the region is illustrated in Figure 8. The maximum difference of 1.08° C occurred at U of W Station 29, about 120 miles south and east of Kodiak Island. Other observed large differences were 0.93° C at U of W Station 10, located about 100 miles southwest of Yakutat Bay; 0.91° C at U of W Stations 6 and 8; and 0.81° C at POG Station 14, located west of the Queen Charlotte Islands. The difference was less than 0.2° C at the coast except off Vancouver Island and Cape Flattery where it was greater than 0.5° C. Also in the south central part of the region the difference was less than 0.2° C. The average value of the observed differences is 0.33° C. The variation of differences between the maximum and minimum temperatures was too large to conclude that the vertical

structure is the result of only vernal heating and wind mixing. The several isolated large differences must have been due to some process such as upwelling or shearing in the water column. Similarly, the absence of a dicothermal structure must also have been a result of motion in the water column. It is concluded that large differences between dico- and meso-temperatures (differences of the order of 1 C°) and the absence of a dicothermal temperature are both non*characteristic.

Below the relative maximum temperature a smooth temperature decrease with depth occurs, at least to 3900 meters as indicated by the temperatures of U of W Station 55.

Vertical sections of temperature. Figures 9a through 9f illustrate the distribution of temperature in the six selected sections.

All sections had a well-delimited thermocline which occurred at 25 to 50 meters depth. The minimum temperature at the top of the thermocline was about 10° C for most sections, but was higher along the eastern and southern limits of the area. At the southern side of the region the temperature at the top of the thermocline reached 14° C, and 100 miles off Cape Flattery was as much as 16° C.

The temperature at the bottom of the thermocline was a minimum in the center and at the western side of the region.

In the sections along Longitudes 161° W and 147° W, the isotherms below the thermocline rose from south to north until some shoalest depth was reached within 200 miles of the edge of the continental shelf. Closer to, and at the shelf, the isotherms descended rapidly. This is best exemplified by the 4° C isotherm in both sections. In a south-to-north traverse of the sections, the 4° C isotherm rose and then fell at least

350 meters.

The rising and falling of isotherms below the thermocline did not occur in Section 135. Furthermore, nowhere in Section 135 was the temperature less than 4°C above 500 meters. Thus the water represented in this Section is about 1°C warmer than the other meridional sections.

The section along Latitude 54°N had a similar temperature structure to Sections 161 and 147, except that adjacent to the continental slope the isotherms below the thermocline fell to seaward. However, further east the 4°C isotherm rose about 350 meters. From mid-section, temperature at any depth increased eastward.

In Sections 51 and 47 the rising and falling of isotherms did not occur (except for meanderings caused by the dico- and mesothermal structure). Instead the temperature at any depth generally increased from west to east. This accounted for Section 135 being the warmest meridional section.

At the east end of Sections 51 and 47 the thermocline rose and, in the case of the latter section, intersected the surface. Indications are that a few miles east of U of W Station 69 of Section 51 the thermocline also reached the surface. The upward slope of the thermocline toward the Canadian coast has been discussed by Doe (1955). It was observed in 1950 and in 1951 and it occurred in the vicinity of the continental slope.

Lateral Distributions of Temperature. The horizontal temperature distributions at various depths are shown in Figures 10a through 10h. At each depth, isotherms were essentially zonal south of Latitude 50°N

on the western portion of the region. Further east, isotherms turned north and then ran approximately parallel to the coast line.

The temperature distribution at 10 meters depth was of marked zonal character south of Latitude 50° N and between Longitudes 160° W and 140° W. The southern limit of the region had temperatures greater than 14° C. Still higher temperatures occurred in the southeastern part of the region, with the maximum observed temperature, 16.37° C, located about 75 miles southwest of Cape Flattery. The 11° C isotherm had the core shape, but south of the Shumagin Islands the isotherm turned north and east and ended at about Latitude 54° N. A tongue of relatively high temperature water was present in the northern part of the region. From the warm tongue to the coast of the Alaska Peninsula temperatures decreased. The minimum observed temperature, 8.62° C, was located near the Shumagin Islands. Relative minimum temperatures also occurred off Cook Inlet, Dixon Entrance, and along the Canadian coast from Queen Charlotte Sound at least to Cape Flattery.

On the 50 meter surface, temperature was less than 5° C at the western edge of the region. East and south of Kodiak Island there was an area with temperature slightly less than 5° C. In general the core area had temperatures between 5° C and 6° C. From the core area to the Alaska and Kenai Peninsulas higher temperatures occurred, in contrast to the 10 meter distribution. Only off Vancouver Island did relative minimum temperatures exist. Two small areas of relative maximum temperature were present. One was located near the center of the region and had a maximum observed temperature of 9.80° C; the other was located south and west of Cape Flattery, within the 9° C isotherm tongue, and

had the maximum temperature of 11.79°C . Southeast of the latter area the temperature was less than 8°C , and northeast (or off the coast of Vancouver Island) was less than 7°C . On the southwest corner of the region temperatures greater than 7°C occurred.

At 100 meters depth the core pattern was well expressed by the 4°C and 5°C isotherms. In the northeast part of the region the 5°C isotherm curved to form a tongue of relatively high temperature water, centered on U of W Station 6. There the temperature was 5.98°C while at the neighboring U of W Station 7 a value of 4.53°C occurred. North, east, and south of the core area temperature increased with maxima, greater than 7° , occurring at the southeastern limit of the region and next to the Queen Charlotte Islands. The maximum observed temperature, 7.99°C , occurred at U of W Station 64, located southwest of the Queen Charlotte Islands, and this produced an area of relative maximum temperature. The minimum observed temperature at 100 meters depth was 3.74°C and was located in the core area.

At 200 meters depth the isotherm pattern did not represent the core pattern as well as at 100 meters. In general, temperatures in the core area were greater than at 100 meters depth. A tongue of temperature slightly less than 4°C extended from the western boundary of the region to about Longitude 154°W . In the center of the region there was an area of temperature less than 4°C . Here the minimum observed value was 3.85°C . The 5°C isotherm was essentially zonal along the southern side of the region, as far as Longitude 140°W where it turned north. At Latitude 55°N the 5°C isotherm turned west and south and back again to form a tongue of relatively warm water, in the same manner as at 100

meters. At the center station of this tongue, U of W Station 6, temperature was 5.61°C at 200 meters depth, while at the neighboring U of W Station 7 the value was 3.92°C . East and south of the 5°C isotherm the temperature generally increased, with maximum values near 7°C occurring at the southeast corner of the region. Two areas of relative maximum temperature existed. One was located at U of W Station 63, where the observed value of 5.64°C was surrounded by observations of less than 5°C . The other included POG Station 13 and was located due west of the Queen Charlotte Islands. Here a temperature of 6.40°C was surrounded by values less than 6°C .

At 400 meters depth temperatures were less than 4.0°C in most of the central and western portions of the region. The minimum observed value was 3.61°C . The 4.0°C isotherm, which turned north at about Longitude 143°W , swung east at Latitude 55°N and back again further north, thus forming a tongue of relatively cold water. The tongue included U of W Station 7 which had the temperature 3.78°C . In contrast, U of W Station 6 had a value of 4.60°C , and this showed as a relative maximum bounded by the 4.5°C isotherm. At the southern limit of the region, the 4.5°C isotherm was near-latitudinal to about Longitude 135°W . Then this isotherm turned north and curved to and from the eastern coast to form a series of tongues. At Latitude 55°N the isotherm approached the coast. The tongue best defined by the 4.5°C isotherm included, at the tip, U of W Station 63 and, mid-way, POG Station 13, the latter had the relative maximum temperature of 5.16°C . South of that tongue was a small area with temperature slightly less than 4.0°C . South and east of the 4.5°C isotherm, the temperature increased. The

maximum observed value was 5.48°C and was located about 50 miles southwest of Cape Flattery.

At 600 meters the core area had temperatures less than 3.5°C . The minimum observed value was 3.27°C . The 3.5°C isotherm was extended to a tongue on the northeast to include U of W Station 7, where the temperature was 3.46°C . At the neighboring U of W Station 6 the relative maximum value of 4.00°C occurred. On the southern limit of the region the 4.00°C isotherm began at Longitude 134°W and ran toward Vancouver Island, then curved from and to the coast, almost meeting the coast at Latitude 55°N . The last turning formed a tongue of relatively warm water and included there was POG Station 13 at which the temperature was 4.25°C . East of the 4.0°C isotherm the temperature increased. The maximum observed value was 4.55°C and was located about 50 miles southwest of Cape Flattery.

At 1000 meters depth, the core pattern was expressed by the 2.75°C isotherm and the 3.0°C isotherm. At Latitude 55°N the 3.0°C isotherm formed a relatively warm tongue that extended southwest to include U of W Station 63. Further north the 3.0°C isotherm again curved, there to include U of W Station 6. POG Station 13 had the relatively high temperature of 3.25°C . At the eastern boundary of the region, from Dixon Entrance southward, temperatures higher than 3.25°C existed. The maximum observed temperature was 3.50°C and was located about 50 miles southwest of Cape Flattery. The minimum observed temperature was 2.62°C and was found in the core area. Thus at 1000 meters depth the observed range in temperature was appreciable.

The observed temperature range at 2000 meters was much less than

at 1000 meters. The minimum was 1.80° C and the maximum was 1.96° C. Lowest temperatures (less than 1.90° C) were found in a band delimited roughly by Latitudes 51° N and 54° N. A tongue of water of temperature less than 1.90° C extended northeast from Longitude 148° W, running almost to the head of the Gulf of Alaska.

THE SALINITY FIELD

Characteristics of the vertical distribution of salinity. A typical example of the distribution of salinity with depth is illustrated in Figure 11. The curve exhibits an essentially isohaline surface layer, an intermediate zone in which salinity increases rapidly with depth (halocline), and a bottom or lower zone in which salinity increases slowly with depth.

In what follows, the characteristics of the halocline zone are discussed in some detail. It should be noted that in such a discussion of vertical salinity structure there is nothing to supplement the data given by the serial observations, i.e., there is nothing which corresponds to a bathythermogram. For this reason a very detailed investigation of the salinity structure of a water column cannot be given.

The depth of the top of the halocline (or of the bottom of the surface zone) varied from 55 meters to 140 meters with the mean of 95 meters. However, the variability in depth had no geographical significance but was quite random, such that large differences occurred between adjacent stations. This was not the case for the salinity at the top of the halocline. Figure 12 shows that this salinity was least at the coast and was highest in the core area, and at the southwest limit of the region. The observed minimum and maximum values of this salinity are 32.47 ‰ and 33.20 ‰ respectively.

Tully (1953) found that the bottom of the halocline was indicated by the salinity $33.8\text{ ‰} \pm 0.1\text{ ‰}$. Thus the topography of the bottom of the halocline is obtained by contouring the 33.8 ‰ isohaline surface. This has been done in Figure 13, Shoalest depths occurred in the core

area, and off Cape Flattery. The minimum observed depth in the core area was 117 meters. North of the core area the depth of the index salinity increased quickly, exceeding 200 meters just south of Kodiak Island. Southeast of the core area a slow increase in the depth of the 33.80 ‰ isohaline occurred with depth exceeding 200 meters in a band which ran approximately southwest from the Queen Charlotte Islands. Within the band, depths exceeded 225 meters in two small areas and at POG Station 13 the index salinity was at 288 meters. East of the band of maximum depth, the index salinity depth decreased uniformly to 90 meters just off Cape Flattery.

Vertical sections of salinity. The salinity structure of the six vertical sections are illustrated in Figures 14a through 14f.

The halocline is readily discernible in all sections and is delimited by the isohalines 33.00 ‰ to 33.75 ‰ (33.80 ‰). The sections along Longitudes 161 and 147 show that the halocline rose from the south, reaching shoalest depths in the core area, and descended near the continental shelf of the Alaska Peninsula. Near the Canadian coast the halocline rose.

Maximum surface salinity occurred in the center of the region and lowest salinity values were found near the coasts. Adjacent to the Alaska Peninsula, isohalines sloped down toward the coast, exhibiting a typical salt-wedge pattern. Near the Canadian coast the opposite was true. Here isohalines sloped upward toward the coast, so that the halocline formed a trough off the coast. This structure is reported by Doe (1955).

Lateral distributions of salinity. Figures 15a through 15h show the distribution of salinity at depths 10, 50, 100, 200, 400, 600, 1000, and 2000 meters.

At depth 10 meters the maximum salinity values (greater than 32.75 ‰) occurred in the center of the region, while minimum values occurred at the coast. West, north and east of the area of the maximum, the salinity decreased, but southward the salinity first decreased and then increased. At the southern limit of the region the essentially latitudinal isohaline of 32.75 ‰ intercepted the surface, and south of this the surface salinity continued to increase. The lowest observed value of salinity was 30.07 ‰ and was located in Shelikof Strait, between Kodiak Island and the Alaska Peninsula.

At 50 meters depth the salinity was between 32.75 ‰ and 33.00 ‰ in half the region. Toward the coast lower values are found. Minima of less than 31.75 ‰ occur in the vicinity of the Shumagin Islands and at the mouth of Cook Inlet. From the center of the region to the coast of Vancouver Island, salinity first decreases slowly to less than 32.50 ‰ and then increased quickly to greater than 33.00 ‰, the observed maximum of 33.18 ‰ occurring just off Cape Flattery. At the western edge of the region salinity values of less than 32.75 ‰ existed. As at 10 meters, salinity increased southward and the 33.00 ‰ isohaline occurred on the western half of the southern edge of the region.

At 100 meters depth the isohalines defined a large area of salinity between 32.75 ‰ and 33.00 ‰. This area extended as a band about 250 miles wide from the western limit to just over half way across the region, at which point the band spread north and south. South and east

of this area salinity increased. Northwestward the salinity first increased to the observed maximum of 33.77 ‰ about 140 miles east of Kodiak Island, then decreased toward the coast. The minimum observed salinity was 31.76 ‰ and occurred between the Shumagin Islands and the Alaska Peninsula. Isohalines north of the band area approximated the core pattern exhibited by the temperature isotherms.

Whereas at 100 meters the observed difference between maximum and minimum salinities was 2.01 ‰, at 200 meters this difference was 0.46 ‰. Most of the region had salinity between 33.80 ‰ and 34.00 ‰. Salinity was greater than 34.00 ‰ in a narrow tongue extending from the western edge of the region to Longitude 150° W, along the axis of the core area. Minimum salinities (less than 33.80 ‰) were found in a tongue extending southwest from the Queen Charlotte Islands to the southern limit of the region, and in two small areas in the southeast part of the region. Adjacent to Vancouver Island and the Washington coast salinity values were greater than 33.90 ‰. This was a secondary maximum.

At 400 meters and at all greater depths to 2000 meters the lateral salinity distribution had the typical core pattern exhibited by the corresponding temperature distributions. Maximum salinities were found in the northwestern part of the region.

Though at 1000 meters the observed range in salinity was only 34.31 ‰ to 34.44 ‰, that at 2000 meters was even less, i.e., 34.57 ‰ to 34.65 ‰. In spite of the small range, the core pattern was evident at 2000 meters. Highest salinities occurred at the southeastern corner of the region.

THE MASS FIELD

Characteristics of the vertical density structure. An example, typical of the distribution with depth of density anomaly (σ_T)¹ is shown in Figure 16. From the surface down, the curve exhibits: a shallow surface layer in which σ_T is constant; a shallow zone in which it increases rapidly; a zone in which it increases slowly; another but thicker zone in which it increases quickly; a lower zone which is characterized by a slow increase with depth.

The shallow constant-density surface zone is concurrent with isohaline and isothermal conditions. The initial severe temperature decrease of the thermocline induces the shallow layer of marked density increase. This is termed the "temperature-dependent pycnocline". Small temperature variations and near-isohaline conditions account for the intermediate zone of small density increase. The second zone of marked density increase occurs within the halocline and so is termed the "salinity-dependent pycnocline". The small density increase with depth in the lower zone is the result of similar gradients of temperature and salinity.

Vertical sections of density. The distribution of σ_T in the six selected vertical sections are shown in Figures 17a to 17f. In general each section shows the double pycnocline; the first occurred at the depth of the thermocline, and the second and less well-defined occurred in the halocline.

¹ $\sigma_T = (\text{density}_{S,T,0} - 1) \times 1000$ expressed as milligrams per cm.³.

In the section along Longitude 161 the double pycnocline structure is clearly marked. Surface σ_t values were highest in mid-section. In general the isopycnals rose slightly from the south end of the section and then dropped abruptly at the edge of the continental shelf. The salinity-dependent pycnocline rose about 25 meters at a distance 100 miles south of the shelf edge and the double nature of the density structure was lost there. The 27.00 isopycnal rose from 410 meters at the south end of the section, to 195 meters about 60 miles south of the continental slope, then descended to 350 meters at the slope.

In the section on Longitude 147 the double pycnocline structure was not as well marked as in Section 161. Surface σ_t values increased from the south to a maximum at mid-section, then decreased northward, becoming very low (< 23.25) at the northern coastal end. As in Section 161, surface layer isopycnals rose slightly from south to north until over the edge of the continental shelf. There isopycnals dropped sharply. The 26.00 isopycnal occurred at 100 meters depth south of mid-section and rose northward to 40 meters. The 26.25 and 26.50 isopycnals showed a similar trend. At 285 meters depth the 26.75 isopycnal entered from the south, rising to 115 meters 100 miles from, and dropping to 170 meters at the continental shelf. The 27.00 isopycnal was much the same, rising from 435 meters at the southern limit of the region to the minimum 225 meters about 200 miles south of the continental slope, after which it dropped to 325 meters at the slope.

Again in the section along Longitude 135 the double pycnocline structure was not well marked but it was present. The salinity-dependent pycnocline was essentially level along the section. The mean density of

a water column was less than in Section 147. The vertical density structure was very much like that of the south end of Section 147.

Surface σ_T values of the section along Latitude 54 exceeded 25.00 in the middle of the section and became less than 24.25 at the east end. West of mid-section, the surface values decreased to less than 24.50 at POG Station 36, and rose slightly to 24.75 at the west end. All isopycnals which were continuous across Section 54 had a shoalest depth in mid-section and in general descended toward the east end. West of mid-section the isopycnals descended until about 125 miles off the continental slope, then rose slightly to the slope.

In the section along Latitude 51° N the surface σ_T values were less than 25.00 at the west end, greater than 25.00 close to mid-section, and became less than 24.25 at the east end, over the continental shelf at Queen Charlotte Sound. The temperature-dependent pycnocline contained the isopycnals 25.25 to 25.50 throughout the section. It was sharply delimited on the west side of the section. Eastward, with the surface waters becoming less dense, lesser isopycnals entered this pycnocline, and some of those which defined the pycnocline on the west diverged to form a smooth density gradient. The salinity-dependent pycnocline was defined throughout by the isopycnals of 26.25 to 26.50 and was supported on the west by that of 26.75, and on the east, by 26.00. In general then the salinity-dependent pycnocline became weaker from west to east. This pycnocline reached a maximum depth at U of W Station 64 (about 300 miles west of the continental shelf edge) and was 50 meters shallower over the shelf edge. In contrast, the 26.75 isopycnal descended about 50 meters over the same distance, and all deeper isopycnals in general

descended from west to east. The rising of the halocline-dependent isopycnals and the descent of deeper isopycnals near the Canadian coast has been observed by Doe (1955).

The σ_T distribution in Section 47 was essentially the same as that of Section 51. Differences between the two sections were due mainly to the southern section being the warmer. Higher surface temperatures resulted in lower surface σ_T values and a well-delimited temperature-dependent pycnocline. Isopycnals from 26.00 to 26.75 again descended from west to east, reaching a maximum depth at POG Station 81, about 400 miles southwest of Cape Flattery. Eastward those isopycnals rose sharply toward the coast. The 27.00 isopycnal fell about 100 meters in its traverse of the section.

Lateral distributions of density. Figures 18a through 18g show the lateral distribution of σ_T on those surfaces for which temperature and salinity were previously indicated.

The distribution at 10 meters was characterized by maximum density in the center of the region - not in the core area. In all directions from the high density center, lower σ_T values were found. Southeast of the Kenai Peninsula a small area of secondary maximum density occurred.

At 50 meters the center of maximum density was larger and was shifted northwest from the center of the region. Two other areas of similar maximum density occurred; one extended as a tongue into the west side of the region, the other lay off Cape Flattery. The latter concurs with the vertical Section 47. Northwest of the center of maximum density a marked gradient exists. South of the high density center there was a

small area of relative low density. Two tongues of relatively low density water were at the southeast corner of the region, one of which imparted a marked gradient in the vicinity of Cape Flattery.

One hundred meters was the shoalest depth at which a lateral σ_T distribution exhibited the typical core pattern. From the core outwards, density decreased except for two areas which were relative maxima; one was on the southwest corner of the region and the other was off Vancouver Island and the Washington coast.

At 200 meters and at all depths to 1000 meters, the lateral density distributions had the core pattern, but lateral density gradients decreased with depth.

SUMMARY AND CONCLUSIONS

The vertical structure of temperature is characterized by

1. a surface isothermal layer of average depth 23 meters which is presumably due to wind mixing;
2. a zone in which temperature decreases rapidly with depth (thermocline) to a relative minimum temperature. The average temperature decrease was 6.9°C ;
3. a relative minimum temperature and the corresponding relative maximum temperature below it. The average difference between the two temperatures was 0.33°C . Both large differences (about 1°C) and the absence of diathermal structure are considered non-characteristic;
4. a smooth temperature decrease with depth below the temperature maximum.

Below 100 meters depth the lowest temperatures at any level were in the northwestern part, or core area, of the region. There was a strong temperature gradient between the core area and the Alaska Peninsula and similar but weaker gradients east and south of the core. Maximum temperatures occurred off Cape Flattery.

The vertical structure of salinity is characterized by

1. an essentially isohaline surface zone;
2. a zone in which salinity increases rapidly with depth (halocline) until the salinity 33.8% is reached. The halocline was essentially horizontal in the offshore region but it was slightly domed in the core area. In the vicinity of the continental shelf the

halocline rose off the Canadian coast and descended off the Alaska Peninsula;

3. a lower zone in which salinity increases slowly and evenly with depth.

At 200 meters and deeper, the lateral distributions of salinity exhibited the core pattern, with highest salinities in the core area. As for the temperature field, strongest salinity gradients occurred northwest of the core area.

The vertical structure of density (σ_T) was characterized by two pycnoclines. The shallower and sharper density increase was related to the thermocline and was denoted the temperature-dependent pycnocline. The other deeper and weaker pycnocline was associated with the halocline and is termed the salinity-dependent pycnocline.

Below 200 meters depth the computed σ_T lateral distributions were, of course, the same core pattern distributions as salinity and temperature.

The existence of the dico- and meso-thermal temperature structure and of the halocline suggests a surface influence on structure. The influence extends below 100 meters but in general not below 200 meters. Therefore the bottom of the halocline is a good approximation of the bottom of the ocean troposphere. In turn, 200 meters is a good approximation of the bottom of the halocline.

The six vertical sections suggest that below the limiting depth of the surface zone (i.e. below 200 meters) the shape of any surface of constant property must be that of a skewed dome. This is verified by Figures 19, 20 and 21 which illustrate the topography of the 4° C

isothermal surface, the 34.0 ‰ isohaline surface, and the 27.25 σ_t isopycnal surface. In each case the shoalest depth of a surface was in the core area. Away from the core area depth of these surfaces increased, but the declivity of each surface was greatest north of the core area. Maximum depths of the surfaces were found adjacent to the Canadian coast except in the case of the salinity surface, which was deepest about 300 miles west of Cape Flattery.

It is to be noted that the observed range in depth of the 34.0 ‰ isohaline surface (190 meters to 485 meters) fell well within that of the 4° C isothermal surface (55 meters to 780 meters). This indicates that the isohaline surface was flatter than the isothermal surface. Thus there was an isotherm pattern on the σ_t surface with lowest temperatures in the core area. This suggests that the domed shape of a σ_t surface is not caused only by the uprising of water in the core area.

From the "skewed dome" model, it is possible to envisage a circulation for the region. Current speed may be assumed approximately proportional to the inclination of the σ_t surfaces and current direction to be along contour lines with greater depths on the right of the current. Figure 21 suggests that there was a slow latitudinal movement eastward across the southwest part of the region. This veered north at mid-region to circulate around the Gulf of Alaska. The great inclination of the surfaces north of the core area indicates that there was an intense stream to the southwest along the Alaska Peninsula.

CHAPTER II

CURRENTS AND VOLUME TRANSPORT

In what follows, the presumed equipotential surface of geostrophic calculations is established at 2000 decibars. Next a new method is suggested for overcoming the difficulty presented by the extension of the level of no motion into the sea bottom. Then follows a discussion of the distribution with depth of current speed, stressing the influence of local surface wind. Examination of velocity profiles near 2000 decibars leads to the postulation of current at great depths, which is further investigated. Then the volume transport of water is discussed in relation to surface wind influence, intensification on the west side of the region, and the effect of bottom topography.

METHOD

The choice of the equipotential surface. Figure 22 illustrates the geopotential topography of the surface of the Northeast Pacific Ocean referred to a presumed equipotential surface at 1000 decibars. The current is presumed to flow along the contours with velocity proportional to the gradient across them. The contours indicate that there was a slow latitudinal movement to the east, south of Latitude 50° N. East of Longitude 145° W the surface current turned north, ran to the head of the Gulf of Alaska and then turned southwest and underwent acceleration. The resulting strong current followed the Alaska Peninsula to Longitude 165° W and west of that it continued along the Aleutian Islands chain and was partially dissipated through the many passes into the Bering Sea.

In the past, the depth of no motion for geostrophic calculations

in this region has been chosen as 1000 decibars. However, it is noted in the present data that the pattern of isopleths of temperature and salinity (and hence of σ_T) at 1000 meters (Figures 10g and 15g) was similar to the contour pattern of dynamic height anomalies referred to a presumed equipotential surface of 1000 decibars. This implied the existence, at 1000 meters, of a current in approximately the same direction as that indicated by the dynamic topography of the upper 1000 meters. Therefore it is logical to presume a deeper equipotential surface.

The distributions of temperature and salinity at 2000 meters are shown in Figures 10h and 15h, respectively. Comparisons with the data from 1000 meters depth shows that there was much less variation of properties, and presumably less geostrophic motion at 2000 meters than at 1000 meters. Hence the 2000 decibar surface was chosen as the equipotential surface.

It is possible that below 2000 meters a level exists at which there is less variation of properties. However, the present data are insufficient to define this, so the choice of the depth of the equipotential surface was limited to 2000 decibars.

The data in Table I are presented as an illustration of the consequences of using a 2000 decibar rather than a 1000 decibar reference level. A comparison of current speeds and volume transports for the two reference levels is shown, using U of W Stations 29 - 30. When the equipotential surface was taken at 2000 decibars, rather than 1000 decibars, there was a current of 4.3 centimeters per second at 1000 meters,

between these two stations. This current was in the same direction as the surface current. The existence of this current at 1000 meters made the surface current larger by a factor of 1.5. The greatest difference, however, occurred in volume transport. In this case volume transport was 2.3 times larger when referred to the lower level.

Dynamic height anomalies. The positions of the stations used for dynamic calculations are shown in Figure 23.

The assumption of 2000 decibars as the depth of no motion precluded consideration of all POG stations and many U of W, and SIO stations, because of either shallow water or shallow sampling. Several U of W stations were sampled to about 1800 meters yet not to 2000 meters. For these, logarithmic extrapolations gave temperature and salinity values at 2000 meters (Tully, 1953). Including these stations, it was possible to calculate dynamic heights referred to 2000 decibars for 36 U of W stations and 6 SIO stations, making a total of 42 stations within the region.

Shoreward of the 2000 meter depth contour of the region there were 5 U of W stations located in water 1100 - 1500 meters deep, which were sampled to about 950 meters. These stations are numbers 2, 12, 23, 52, and 68. It was desirable to use these data for the dynamic assessment of the region, but the calculation of dynamic height referred to 2000 decibars at these peripheral stations implied the extension of the depth of no motion into the sea bottom below the stations.

Several methods are available to overcome the difficulty presented when the depth of no motion enters the bottom. Helland-Hansen

(1934) introduced a method based on the assumption that along the bottom both the horizontal velocity and the slope of the isobaric surface vanish. In effect the curves of specific volume anomaly, δ , are continued horizontally through the bottom from the point of contact. In Figure 24, this is represented by the horizontal broken lines. Jacobsen and Jensen (1926) proposed a similar method but with the additional assumptions that the sea bottom profile is a straight line and that δ curves are parallel and at equal distances. As another approximation, Sverdrup (1942) suggests application of the equation

$$i_p = -\bar{i}_\delta (\delta_1 - \delta_2)$$

where i_p = slope of isobaric surface

$$\bar{i}_\delta = \text{average slope of the surfaces } \delta_1 \text{ and } \delta_2 \\ \text{where they run into the bottom}$$

$$(\delta_1 - \delta_2) = \text{the difference } \delta_1 - \delta_2 \text{ along the bottom.}$$

This is illustrated in Figure 24.

The above methods are, at best, only expedients. All imply knowledge of the slopes of the δ surfaces at the bottom. This can be approximated only through extrapolation of curves. The methods of Helland-Hansen and Sverdrup require knowledge of the form of the bottom profile between stations, while the Jacobsen and Jensen method ignores it. In addition, all the methods require a scaled drawing of the section from which the various slopes and values of δ surfaces may be read off.

Each method may be considered as uncertain.

A new method is now presented. Although it is not more certain, it does not require a scaled drawing of the section or knowledge of the slopes of the δ , or isobaric surfaces. It is necessary to know only the δ values at the observed or standard depths of the shallow or near-shore station and at the reference depth at the closest offshore station. The δ values are plotted against the logarithm of depth. Figure 25 shows the resulting plot for the offshore station, U of W 53. It is a very regular and slightly curved line, at least below the limiting depth of the ocean troposphere (200 meters). The plot for the neighboring near-shore station (U of W 52) is similar but lies below that of the seaward station. The slope of the near-shore station curve is somewhat less than that of the seaward station, particularly at 1000 meters. The reduction in slope at the lowest observed depth of U of W Station 52 suggests that its structure must approach that of the offshore station at greater depth. This, together with the presumption of the 2000 decibar equipotential surface, allows completion of the near-shore station δ curve, with an interpolated smooth curve which approaches the offshore plot roughly asymptotically, meeting at 2000 meters. From the new section of the curve, δ values may be interpolated and computation of dynamic height anomaly proceeds from there. It is stressed that the method is not more certain than any other previously used, but it is as reasonable, and is easier to apply in practice.

Dynamic heights were calculated using the methods described in U.S.N.H.O. Publ. 614 (1951).

Geostrophic current and volume transport. Profiles of mean relative current velocity versus depth were calculated using the formula of Helland-Hansen (1930) and assuming the theorem of circulation of Bjerknes (1933).

$$u_z = \frac{10(\Delta D_A - \Delta D_B)_z}{L_{AB} 2 \Omega \sin \bar{\Psi}}$$

where u_z = relative current at depth z meters between Stations A and B

$(\Delta D_A - \Delta D_B)_z$ = difference at depth z meters between the dynamic height anomalies integrated between 2000 decibars and z decibars for each station.

L_{AB} = distance in meters between stations.

Ω = angular velocity of rotation of the earth.

$\sin \bar{\Psi}$ = average value of the sines of the latitudes of the two stations.

A relative current velocity sign convention is used. A positive current indicates flow with a southwest component; a negative current, a northeast component.

Geostrophic volume transport between stations is computed either through integration of the corresponding velocity-depth profile and multiplication by the distance separating stations, or from the equivalent expression

$$T_z = \frac{10}{2 \Omega} \int_{-2000}^z \frac{[\Delta D_A - \Delta D_B]_z}{\sin \bar{\Psi}} dz$$

where T_z = integrated volume transport between z meters and 2000 meters.

VELOCITY PROFILES

The distributions of mean relative current with depth between each pair of stations may be conveniently classified into four groups. The four groups and the regional occurrence of each are illustrated in Figure 26.

Group I profiles occurred along the western side of the Gulf of Alaska. Current speeds were large and at all depths had a southwest or west component. Maximum current was at or near the surface and was as much as 24 centimeters per second.

Group II are those current speed-depth profiles which at all depths had a northeast current component. Maximum speed was at the surface and did not exceed 5 centimeters per second. Group II profiles occurred immediately seaward of Group I and Group IV profiles, and on the southwestern part of the region.

A surface layer current with a northeast component in contrast to the deeper current characterized the profiles of Group III. Current speeds were small at all depths. The upper current layer ranged in depth from 100 to 300 meters.

Group IV profiles occurred adjacent to the Canadian coast. The relative current was directed northward at all depths. Here maximum velocity occurred at 200 to 400 meters and was about 7 centimeters per second. The surface current was about half the sub-surface maximum current.

Influence of local wind. The profiles of Groups III and IV indicate the effect of some surface acceleration such as wind. This leads to a consideration of the wind field over the Gulf of Alaska. Figure 27 shows the wind roses for the region for each five degree square of latitude and longitude. These data were taken from the Pilot Chart of the North Pacific Ocean for August, 1956. The wind roses represent the average of all winds observed in the month of August for a period of about 100 years. As such, the Pilot Chart indicates the character of expected winds. It is readily seen that winds over the region are relatively weak and variable as compared with, say, the northeast winds that occur at Latitude 25° N. In other words, over the Gulf of Alaska no prevailing winds blow. However, it is possible to construct a vector resultant diagram using wind of constancy 40 to 60 per cent. This is illustrated in Figure 28. It is stressed that the Figure does not represent a distribution of prevailing wind, or a wind pattern, but it is an indication of vectors which exist about 50 per cent of the time, but not necessarily concurrently. This wind tendency is essentially clockwise, except near the head of the Gulf of Alaska. There the wind becomes easterly, but constancy drops to about 20 per cent. The strongest winds (greater than 10 miles per hour) occur in a tongue-like area extending from the west side of the region east to Longitude 135° W and from Latitude 48° N to Latitude 53° N. North, east, and south of the tongue wind strength decreases.

In spite of the lack of a steady wind field over the Gulf of Alaska, there was an evident relation between the surface layer current

and the wind field as such. The most northerly of Group I current speed-depth curves (that of U of W Stations 23 - 24) was located near the head of the Gulf of Alaska where the wind is weak and/or easterly. A strong surface current southwest existed. The most southwesterly profile of Group I (U of W Stations 52 - 53) was in the area of strongest southwest wind. Here the southwest current was a maximum at 50 meters, indicating surface retardation by the wind. Group II and III profiles were found in the central part of the region where the surface layer flow was in the same direction as the wind. Group IV profiles indicate northwest flow against a northwest wind and subsequent surface retardation. The flow indicated by Group IV curves has been observed on "Offshore" surveys (Doe, 1955).

In August, 1955, the wind field, though not strong, influenced the upper zone current pattern. However, the wind did not control the circulation since in some cases the current at all depths was directed opposite to the surface wind. It is concluded that in August, 1955, the general current pattern in the Gulf of Alaska was not related to the local wind system.

THE POSSIBILITY OF ABYSSAL CURRENT

Since the curve of specific volume anomaly plotted against the logarithm of depth always is essentially linear (at least below 200 meters) and since 2000 decibars was chosen as the equipotential surface, then the velocity and velocity gradient should go roughly asymptotically to zero at 2000 meters. All current speed-depth profiles did reach zero at 2000 meters, of course, because this was imposed on the data by the

equipotential assumption. But in some cases there was a change in velocity gradient and a corresponding sharp drop in velocity near 2000 meters. An example is the Group I profile of current between U of W Stations 31 - 30 (Figure 26). From 1000 meters to 1500 meters, the velocity decreased from 2.5 to 2.0 centimeters per second, and in the next 500 meters, decreased to zero. Such a velocity-depth distribution is not impossible, but in this case it appeared that the imposition of the equipotential assumption had precluded current below 2000 meters. If there is current at or below 2000 meters, then the surfaces of the anomaly of specific volume, δ , will not be horizontal at that depth. In Figure 29, the curves of δ versus the logarithm of depth for Stations 30 and 31 are shown. The curve of Station 31 lies below that of Station 30, and the two are nearly coincident between 1200 meters and 1500 meters. But below 1500 meters the curves diverge, at least to 2000 meters. Therefore the δ surfaces could not be horizontal in the vicinity of 2000 meters.

It is now expedient to investigate the distribution with depth of δ surfaces near 2000 meters. Figure 30 illustrates the δ field at that depth. Around the periphery of the Gulf of Alaska at 2000 meters and in the center of the region, δ values were greater than 50 (for convenience, the factor-unit combination 10^{-5} cc/gram is omitted), the maximum value observed being 52.5. The core area and the southwest and southeast corners of the region had δ values less than 50. A minimum observed value of 48.2 was located in the core area, while in the southeast corner of the region there was a minimum of 47.4. It is apparent that in the vicinity of the core area the δ pattern at 2000 meters was

essentially the same as the surface current indicated by dynamic heights referred to 1000 decibars (cf. Figure 22). Then if there was current at 2000 meters (current which became zero at some greater depth) there would be circulation around the core area in the same direction as the indicated surface current.

The broken lines in Figure 30 indicate where the vertical sections of Figure 31 are located. Section I runs southwest from Kodiak Island to Latitude 55° N and then due east to Dixon Entrance. Section II is along Latitude 51° N from Longitude 161° W to the entrance to Queen Charlotte Sound. The sections show the distribution with depth of the specific volume anomaly, from 800 meters depth to 2400 meters (where observations exceeded 2000 meters).

In Section I the δ isopleth 70 entered on the west at 1190 meters, rose to 940 meters at Station 28, and fell to 1150 meters at the east end. The isopleth of 60 was at 1580 meters at Station 31, 1250 meters at Station 28, and 1500 meters at Station 3. The sharp slope of the δ surfaces 70 and 60 at the west end of the section was not reflected as much by the 50 isopleth. However, the latter rose from 2120 meters at Station 31 to 1930 meters at Station 28. Eastward the isopleth dropped to 2000 meters in mid-section, rose to 1920 meters at Station 5, and dropped to 2020 meters at the east end of the section. The data gave four depths of the 48 isopleth. The range of those depths was 2070 meters to 2260 meters. In Section I, therefore, the isopleths of specific volume anomaly in the vicinity of 2000 meters reflected, at least in part, the undulations of those δ surfaces at shoaler depths. The isosteres at 2000 meters are not horizontal.

In Section II, the δ isopleths of value 70 and 60 descended from west to east. The 70 isopleth was at 940 meters at Station 55, rose slightly to 910 meters at Station 59, and then dropped to 1150 meters at the east end of the section. The 60 isopleth occurred at 1300 meters at the west end and 1540 meters near the east end. The 50 isopleth fell from 1880 meters at Station 56 to 2160 meters at Station 58. Eastward the isopleth undulated between this and 2070 meters at Station 67. The data gave seven depths and indicated another of the 48 isopleth. On the western portion of the section the isopleth fell from 2250 meters to more than 2500 meters at Station 59. Eastward the isopleth rose steadily to 2300 meters at Station 67. Then over the eastern half of the section the 48 isopleth assumed a slope contrary to that of the 70 and 60 isopleths. Therefore, down as far as 2500 meters, the δ surfaces do not become and remain horizontal.

The slope of the 48 isostere over the eastern half of Section II appears definite. If all deeper isosteres have the same slope, then there would be a southward current below a depth of no motion near 1000 meters. It follows that, because data from great depths are lacking, no guess can be made of the direction of flow of the inferred abyssal current. Similarly, no guess can be made of the best equipotential surface, so its definition must remain as the level of least lateral gradients of properties, within the range of the data.

It is concluded that motion exists at least to 2000 meters, and possibly below that level, but that the direction of abyssal current cannot be determined with the present data.

The assumption of no motion at 2000 meters resulted in a mean

northward current of 0.15 centimeters per second at 1000 meters in Section II. Though the current speed was small, the difference in transport induced by precluding motion between 1000 and 2000 meters would have been 4.21×10^6 cubic meters per second or about 44 per cent of the transport relative to 2000 meters. Suppose now that at 2000 meters a mean current of only 0.05 centimeters per second existed across the section, and that it was directed north, as is the surface current. If this current went linearly to zero at 2500 meters (an arbitrary choice) then the assumption of zero motion at 2000 meters would not account for a transport of $2.39 \times 10^6 \text{ m}^3/\text{sec.}$ through Section II. Thus a current of 0.05 centimeters per second at 2000 meters with zero motion at 2500 meters would make the volume transport 25 per cent larger than through the same section with motion assumed zero at 2000 meters.

This indicates the need for two things. First, observations should be made below 2000 meters in the region. Second, observations of temperature and salinity should be as accurate as possible.

VOLUME TRANSPORT

The equation usually used to calculate volume transport is

$$T_z = \frac{10}{2 \Omega \sin \Psi} \int_{-2000}^z [\Delta D_A - \Delta D_B]_z dz$$

The equation is correct for latitudinal flow, i.e., true geostrophic flow. However, longitudinal flow is not geostrophic flow. Since the currents in the region are primarily longitudinal, the equation as such cannot be used.

If it is assumed that absolute vorticity is conserved by the flow in the region, then the equation may be used with a slight modification.

The absolute vorticity of a water particle is

$$\xi_0 = 2 \Omega \sin \Psi + \xi$$

where $2 \Omega \sin \Psi$ is the vorticity due to earth's rotation (Coriolis force) and ξ is the vorticity relative to the earth. Now when flow changes latitude there will be a change in Coriolis force. If absolute vorticity is conserved, there must be an equal and opposite change in relative vorticity. Therefore, by introducing the term for relative vorticity in the geostrophic volume transport equation, the equation can be used where the absolute vorticity of flow is conserved. The equation now is

$$T_z = \frac{10}{\text{constant}} \int_{-2000}^z (D_A - D_B) dz$$

The constant should be the sine of the latitude at which the relative vorticity is zero. Thus, if somewhere in the region latitudinal flow existed, then a good choice for the constant would be the sine of the mean latitude of the latitudinal flow. However, within the region there is no marked latitudinal flow. For this reason, the constant was chosen as the sine of the mean latitude of the region, Latitude 53° N.

For the region of investigation the greatest difference between a transport calculated assuming conservation of absolute vorticity and the corresponding geostrophic transport is about 10.5 per cent.

The vector resultant of volume transport integrated over 2000 decibars for each station pair is shown in Figure 32. Volume transport in the southeast corner of the region was the upper half of a large

anti-clockwise eddy of radius about 300 miles. Flow in the eddy was about 3 million cubic meters of water per second. North of the anti-clockwise eddy was a smaller but more intense clockwise eddy whose flow comprised about 8.5 million cubic meters of water per second. Between this intense eddy and an apparent strong northward flowing current near the Queen Charlotte and Vancouver Islands lay a weak elongated anti-clockwise eddy. The northerly near-shore current comprised at least 3 million cubic meters of water per second. The term "at least" must be employed because observations did not extend to the coast. West of the eddies, or in the center of the region, the mean current flow was north to northeast. The boundary between this flow and the opposite flowing eddy of the southeast corner ran approximately west and then southwest from Vancouver Island. The broad northerly current in the center of the region had the magnitude of 14 million cubic meters per second. At Latitude 51° N the flow curved to the right and then back again. A similar but more marked deviation occurred at Latitude 55° N. The deviation of the right half of the flow had the form of a "hairpin" curvature. A small cyclonic eddy was located in the eye of the "hairpin". As recovery from the sharp deviation occurred, the near-shore current of the North American coast added to the transport and the whole continued northwest around the head of the Gulf of Alaska. West of Longitude 145° W the flow turned southwest and underwent acceleration with the result that marked intensification occurred. A stream 100 miles wide which passed at least 17.5 million cubic meters of water per second coursed southwest following the Alaska Peninsula. At Longitude

155° W the stream turned west and carried out of the region. This strong current may be called the "Alaskan Stream."

Influence of surface wind. A comparison of the 0 - 2000 decibar volume transport pattern with that of the summer surface wind field (Figure 28) shows a marked difference. Over the large cyclonic eddy and near the Canadian coast the wind usually blows in directions opposite to that of the transport.

Since the limiting depth of surface influence was defined as 200 meters, and since it was established that there was a wind effect on upper zone currents (page 36) then the 0 - 200 decibar volume transport, relative to 200 decibars, is the integral of surface wind influence.

The vector resultant of volume transport of the ocean troposphere, relative to 200 decibars, is shown in Figure 33a. Comparison with the surface wind distribution (Figure 28) shows the almost complete correlation of wind-field and transport. Only adjacent to the Alaska Peninsula was the upper zone current, relative to 200 decibars, directed up-wind. In this area only the near-surface 50 meters of water was retarded (Figure 26), and integration from 200 decibars masked that feature.

The most marked differences between the flow patterns of Figure 32 and Figure 33a are on the east side of the region. Off the Canadian coast, a southeast flow was indicated when the reference level is 200 decibars, but the flow was northwest when referred to 2000 decibars. Over the southeast corner of the region the circulation was essentially clockwise for the shallow reference surface and anti-clockwise for the 2000 decibar equipotential surface.

It is stressed that the upper zone transport pattern of Figure 33a is relative to 200 decibars and as such is not the "absolute" transport of the upper zone. Figure 33b shows the transport pattern of the same upper zone but referred to the 2000 decibar surface. This flow pattern is the "absolute" circulation of the upper 200 meters. The magnitude of the flow was about five times larger than that of the "relative" flow. Off the Canadian coast the flow indicated was northerly. Over the southeast corner of the region the current was essentially latitudinal.

It is concluded that in August, 1955, in this region of the Northeast Pacific Ocean the surface wind influenced but did not control the surface layer circulation, and further that deep circulation existed independent of the local wind field.

Comparison with conditions in August, 1950. The intense anti-cyclonic eddy of Figures 32, 33a and 33b was observed by Doe (1955) during August, 1950. Figure 34 shows that at that time the eddy was centered about 100 miles southwest of its position in August, 1955. Furthermore, in 1950, an elongated anti-clockwise eddy lay east of, and adjacent to the intense coastal eddy, and the near-shore current off the Queen Charlotte Islands was northerly. These last two recurred in 1955 (Figure 32). In the southern part of the region, in 1950, the flow indicated was essentially latitudinal, except for the southeast corner where a cyclonic circulation existed. This is much like the flow pattern of Figure 33a. It is concluded that, within 600 miles of the Canadian coast, the circulation pattern which existed during August, 1950, was similar to that of August, 1955.

The eddy system was not present in August, 1951. Therefore it appears that this feature is recurrent, but not necessarily perennial.

Vertical motion at the null flow line. At the boundary between the slow cyclonic eddy and the northerly current in mid-region (Figure 32) it is expected that there should have been, somewhere in the water column, either or both of a convergence or divergence. From Figure 33b it is known that the mean transport of water above 200 decibars was opposite to the flow of the cyclonic eddy, or in the same direction as the main surface flow in mid-region. Therefore it is not expected that there was a divergent process at the sea surface along the indicated null line. This was borne out by the lateral distribution of properties at 10 meters depth, for there were no discontinuities in the essentially latitudinal pattern across the boundary. The velocity profiles for the station pairs, in and around the eddy, showed that flow in the direction of the eddy began at roughly 175 meters. Then if an operative divergent process existed, it should have been indicated either by the lateral distribution of properties below 100 meters, or by the topography of surfaces below that depth.

Superposition of the topography of the 33.8 ‰ index salinity (Figure 13) on the 0 - 2000 decibar volume transport pattern shows that along the null line of the flow, the index salinity occurred at a relative minimum depth, never exceeding 205 meters. In each side of the boundary, its depth exceeded 225 meters and attained 245 meters. Thus there was an apparent relation between the depth of the index salinity and the position of the null flow line.

The depth of the temperature minimum and maximum (Figures 6a and 7a) also was a relative minimum along the null flow line. In addition the dicothermal structure was absent in a small elongated area just north of this boundary.

These observations suggest that there was a slight uprising of water at the boundary and a corresponding slight sinking on each side. The process was internal, since it was not indicated at 100 meters and shallower depths.

No estimate can be made of the depth to which the divergent process extends. There are two reasons for this. First, the process is slight, i.e., the slow flows would not result in a marked divergent process. Second, below 200 meters the vertical gradients of the properties are small compared with the gradients in the troposphere. Therefore it would be extremely difficult to observe the vertical motion below 200 meters depth.

It should be noted that Figures 6a, 7a, and 13 include POG data, whereas the volume transport picture does not. Therefore in a discussion of the relation of the topography of the structural features to the figures constitute a semi-independent set of data, and as such impart a greater credence to conclusions than if the structural and dynamic features were based on identical data.

Intensification of Alaskan Stream. The strong Alaskan Stream indicates intensification by some force. If the effect is due to Coriolis force, the Alaskan Stream would be an example of westward intensification. However, this current is not the same as other westward

intensified flows such as the Gulf Stream and the Kuroshio Current. The latter are formed when a broad westward flowing current turns northward. The Alaskan Stream forms after a broad northward flow turns westward and then southwest.

If it is assumed that, at the head of the Gulf of Alaska the current, having turned from north to west, constitutes a westerly flow. Then it is possible to investigate the intensification which Coriolis force would cause. Morgan (1956) showed that a wind-driven westerly current could be turned into an intense southward current if there is no boundary on the south. In this case the northern boundary (coast) will constitute a streamline. In the direction of a southward current there is a decrease in Coriolis force. Since the absolute vorticity of the flow must be maintained constant, there must be an increase in velocity shear across the stream. The water "particles" approaching the coast with larger Coriolis force must acquire a larger positive velocity shear than must those approaching with smaller Coriolis force. Then the intensified stream thus formed will have maximum current at the coast.

The absolute vorticity of flow was defined previously as

$$\xi_a = 2 \Omega \sin \Psi + \xi$$

The relative vorticity of a water element in a streamline is defined as

$$\xi = \frac{\partial v}{\partial n} + \frac{v}{n}$$

where $\frac{\partial v}{\partial n}$ = velocity shear normal to the streamline,

$\frac{v}{n}$ = measure of streamline curvature.

When the radius of curvature of a streamline is large, the term $\frac{v}{n}$ is negligible. Therefore, for straight streamlines,

$$\xi = \frac{\partial v}{\partial n}$$

and
$$\xi_a = 2\Omega \sin\Psi + \frac{\partial v}{\partial n}$$

If absolute vorticity of flow is conserved, then at any point on a streamline

$$\frac{\partial v}{\partial n} + 2\Omega \sin\Psi = \text{constant.}$$

When the flow changes latitude from Ψ_2 to Ψ_1 , where Ψ_1 is the lower, then on a streamline

$$\left[\frac{\partial v}{\partial n} \right]_1 + 2\Omega \sin\Psi_1 = \left[\frac{\partial v}{\partial n} \right]_2 + 2\Omega \sin\Psi_2$$

$$\text{or } \left[\frac{\partial v}{\partial n} \right]_1 - \left[\frac{\partial v}{\partial n} \right]_2 = 2\Omega [\sin\Psi_2 - \sin\Psi_1]$$

If the latitude change is small, then the increase in velocity shear normal to the streamline which must be acquired by the flow when it reaches the lower latitude is

$$\frac{dv}{dn} = 2\Omega \delta\Psi \overline{\cos\Psi}$$

As is indicated by Figure 35, the 0 - 2000 decibar volume transport flow between U of W Stations 23 and 25 formed part of the flow between U of W Stations 45 and 46. Only two of the flow lines are shown in Figure 35. They were essentially straight between the two

station pairs. Therefore, in the direction of flow, the velocity shear normal to the flow line for each flow line must have increased.

The theoretical shear increases normal to the flow lines for the observed latitude changes were calculated. The results are indicated in Table II.

The shear increase normal to the flow lines at AA' varied linearly from 0.40 centimeters per second per kilometer at A to 0.50 centimeters per second per kilometer at A'. The distance between the flow lines at AA' was about 74 kilometers. Integration of the shear increase across that distance gave the surface current profile at AA', due to shear increase between BB' and AA'. This is illustrated graphically in Figure 35.

The mean surface current due to increase in normal shear is about 16 centimeters per second. This is, theoretically, the increase in mean surface current which should have occurred between BB' and AA'. But the observed increase in mean surface current between those two traverses was about 4 centimeters per second.

Thus the theoretical mean surface current increase is four times as large as the observed increase. If it is postulated that friction inhibits intensification by as much as fifty per cent, the theoretical current increase will still be roughly twice the observed increase. The large discrepancy may be due to either of the following:

- a) absolute vorticity of the flow was not conserved and the theoretical calculation is meaningless;
- b) absolute vorticity of the flow was conserved but the theoretical analysis is faulty.

The latter is considered more probable, as the following continuity argument indicates.

The distance across BB' was about 160 kilometers. The mean surface current there was about 9 centimeters per second. The distance across AA' was about 74 kilometers. Then continuity of volume demands that the mean surface current at AA' was about 19 centimeters per second. If this was the case, the increase in mean surface current between BB' and AA' would be about 10 centimeters per second. That would agree much better with the theoretical increase of 16 centimeters per second.

It is suggested that volume transport - not surface current - should be used to assess intensification. Such an analysis involves a third dimension, depth, and becomes rather more than the scope of this paper. However, the simple continuity argument suggests that absolute vorticity of flow is conserved.

Current in the vicinity of Sanak Island. At Longitude 161° W, the northernmost station of depth 2000 meters (extrapolated) was U of W Station 52. Between that station and the southern boundary of the Alaskan Stream the volume transport westward was about 8.6 million cubic meters of water per second. But the circulation in the Gulf of Alaska comprised at least 17.5 million cubic meters of water per second. Then continuity of volume required that at least 8.9 million cubic meters of water pass west per second between Station 52 and the Alaska Peninsula. Again "at least" must be used in discussing transport for here no estimations have been made of contributions to the total circulation by currents along the shores, or by river discharge into the Gulf of Alaska.

It follows, however, that a rough estimation of the mean current between U of W Station 52 and the Alaska Peninsula is possible.

Figure 36 depicts the profile of the bottom between Unimak Island and U of W Station 52. The cross-sectional area of the water in this section is 23.9 million square meters. Then the passage of at least 8.9 million cubic meters of water per second would require a mean current of at least 37 centimeters per second. Such a mean current is not unreasonable, but should be indicated by large slopes of such surfaces as σ_T or δ . That this situation existed is verified by the inset plate of Figure 36 which shows in section the distribution with depth of the δ surfaces between Stations 51, 52, and 53. Station 51 was 27.8 kilometers north of Station 52. The slope of the δ surfaces was much greater between Stations 51 and 52 than between Stations 52 and 53, and it was between the latter pair of stations that a maximum mean current of 24 centimeters per second occurred.

That a strong coastal current exists is verified by the United States Coast Pilot 9 - Alaska, from which the following are quoted:¹

"A continual current of considerable strength follows the coast all the way from Shelikof Strait to the Aleutian Islands."

"The coastal current searches out all passages, large and small, between and around the many islands, and in some of them it becomes strong enough to be important."

¹ United States Department of Commerce, Coast and Geodetic Survey, United States Coast Pilot 9 - Alaska: 1954, (Washington: Government Printing Office, 1955), p. 334.

"On three runs between Chirikof Island and Castle Rock, a survey ship experienced a southerly set indicating an average strength of current of $1\frac{1}{2}$ knots." (75 cm. per second).

Influence of bottom topography. The mean volume transport of water above 2000 meters in the center of the Gulf of Alaska suffered two marked directional changes (Figure 32). The first occurred at Latitude 51° N, where the flow curved first to the right and then back again in its northward travel. Further north and east, at Latitude 54° N, a sharp current deviation to the right occurred and immediately to the north the flow turned west. In the center of the hairpin-like curvature thus formed was a small cyclonic eddy. Such deviations in the direction of volume transport are rather marked to have origin in some non-mechanical process (e.g. thermal factors). Therefore a consideration of a mechanical process leading to current direction changes is necessary for explanation of observations.

Shown in Figure 37 is the bathymetry of the Gulf of Alaska. Superposition of the volume transport pattern on that of the bathymetry clearly indicated that the two changes in direction of the flow occur above chains of seamounts. It is therefore expedient to investigate the influence of bottom topography on the direction of the transport.

Sverdrup (1942) explained topographical effects by taking into account the Coriolis force and the vertical stratification of the water masses. His reasoning is illustrated schematically in Figure 38, a three-dimensional diagram depicting the deformation of the isopycnic

surfaces (ρ) and of the open isobaric surface of the sea which would be caused by a current crossing an underwater ridge. At some distance in front of the transverse ridge, where the bottom is horizontal, the isopycnic and isobaric surfaces are horizontal in the layer adjacent to the bottom, where it is assumed that there is no current. With increasing distance from the bottom the isopycnic surfaces slope, rising from right to left for an observer looking in the direction of the current, while the isobaric and along with them the open surface of the sea, assume a slope contrary to that of the isopycnic surfaces. As the stream approaches the submarine ridge, the isopycnic surfaces in the bottom sub-layer curve up over the ridge. The isopycnic surfaces in the higher layers of the water will also be lifted upwards. This effects a lowering of the isobaric surfaces, and of the sea surface itself, along the course of the current, in such a way that over the ridge there must be a trough-shaped depression in the sea surface. Therefore, as the current approaches the ridge it will deviate more and more to the right and will reach its maximum displacement over the crest of the ridge. Beyond the crest of the ridge a deviation back to the left will occur and at some distance away the current will once again assume its original direction.

Since the deflection is in the same direction at all depths, it is readily seen that mean volume transport integrated from surface to bottom would suffer the same deflection.

In order to apply Sverdrup's theory to explain the direction changes of circulation, an assumption is necessary. The theory concerns the effect on water movement caused by a ridge (or trough), i.e. a

continuous bottom morphological feature. Below the two areas of the Gulf of Alaska where marked changes in the direction of volume transport are observed, there are no ridges or troughs. However in each case a chain of seamounts exists. Therefore it is assumed that a current crossing a transverse chain of seamounts suffers the same change in direction as is effected by a ridge.

With that assumption it is readily apparent that Sverdrup's theory explains the two major current deviations. The short chain of seamounts centered roughly at Latitude 51° N and Longitude 143° W caused the flow to bend to the right as it approached the chain. Over the bottom elevations the maximum deflection occurred and north of the chain the current curved back to the left and then straightened out. The flow deviation at Latitude 55° N was of the same character and was centered on the northern band of seamounts. Therefore it is concluded that, on the basis of Sverdrup's theory, the flow deviations were due to the effect of bottom topography.

Here the assumption is recalled that a chain of seamounts deflects an ocean current in the same manner as a submarine ridge. Since the observed deflections occurred over seamount ranges, and since the deflections concur with theory, the assumption seems reasonable.

A different explanation of the effect of bottom topography on the direction of ocean current has been presented by Shtokman (1948). The controlling factor, along with bottom topographical features, is not the pressure gradient but the wind field. Shtokman argues that if ocean currents are affected by the wind, then the pressure gradient cannot

exist completely independent of the wind. In that case, because of the rotation of the earth the slopes of the isobaric surfaces observed will be the result of an adaptation of the pressure gradient to a Coriolis force set up by an already existing current produced by the wind. It follows that the pressure gradient and density distribution in the depths of the water will be a secondary effect of the wind. Then Shtokman's theory applies to a wind-driven circulation. The current deviations are related to the bottom morphology and vorticity of the wind field.

It has already been indicated that the circulation in the Gulf of Alaska is not wind-driven. It is true that the summer surface wind influences the upper zone currents, but the influence does not overcome the primary circulation pattern. Also, as previously discussed, the summer surface wind distribution exhibits no strong, prevailing winds. These two facts preclude application of Shtokman's theory for explanation of the current deviations.

Nowhere thus far in the discussion of the effect of bottom morphology on currents has the depth of the influencing seamounts been considered. Shown in Figure 39 are the 2000 and 3000 meters contour lines of the seamounts of the northern and southern groups. The solid areas or blackened-in areas indicate where depth is between 2000 and 3000 meters, and the inner stippled areas appearing at some seamounts indicate depth less than 2000 meters. These seamounts occur over those parts of the abyssal plain of the Northeast Pacific Ocean where its depth is 2000 to 2200 fathoms (3700 to 4000 meters). It is obvious that

if a current was limited to 2000 meters, it would not be affected as much by the bottom morphology as a current that extended to 3000 meters or more.

The implication of the last statement is worthy of consideration. The possibility of motion below 2000 meters was inferred in the analysis of specific volume anomaly at and below 2000 meters (see page 37). But no guess of the direction of deep flow could be made. However, the observed bottom topographical effects suggest that the deep current is in essentially the same direction as the surface current, at least in the vicinity of the seamount ranges.

There is yet another implication. If flow in the region is essentially geostrophic flow, then abyssal current would have to be a counter current to the surface circulation. However, if the flow satisfies the conservation of absolute vorticity, then it is not necessary that a deep counter current exists. Therefore, current from surface to bottom could be in the same direction everywhere over the region. It follows that the observed bottom topographical effects suggest that flow in the region conserves absolute vorticity.

SUMMARY AND CONCLUSIONS

The volume transport of water in the Gulf of Alaska was characterized by a slow northeast movement through the center of the region. This flow was about 13 million cubic meters per second. Near the head of the Gulf of Alaska the flow was joined by the transport adjacent to the Canadian coast (4 million cubic meters per second) and the whole turned westward and then southwest, all the while undergoing acceleration. The strong southwest flow is named the Alaskan Stream. In the southeast part of the region, volume transport was the upper half of a cyclonic eddy which comprises about 3 million cubic meters per second. North of the cyclonic eddy was an intense clockwise eddy of 8 million cubic meters per second. Between this eddy and the northerly Canadian coastal current was an elongated cyclonic eddy of one million cubic meters per second.

The volume transport integrated over 2000 decibars had the same direction as the surface wind only in the center of the region. Flow in the upper 200 meters of depth was influenced but not controlled by the local wind field. Thus the circulation in the Gulf of Alaska was not wind-driven locally at the time of these data.

The Alaskan Stream may conserve absolute vorticity, i.e., the intensification may be due only to the changing Coriolis parameter.

The existence of motion below 2000 meters was indicated in the analysis of the data. This view is supported by consideration of the influence of bottom topography on the direction of flow, since the seamount chains of the region would affect a current more if it extended

below 2000 meters. Taken together, these two lead to the conclusion that the inferred current below 2000 meters has the same direction as the observed volume transport of the upper 2000 meters, at least in the vicinity of the seamount chains. This is reasonable if absolute vorticity is conserved at all depths.

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TABLE I

Surface Current Speeds and Volume Transports
between U of W Stations 29 and 30 for a 1000 decibar and a
2000 decibar Equipotential Surface

	Equipotential Level	
	1000 db.	2000 db.
Surface Current (cm./sec.)	9.0	13.3
Volume Transport (10^6 m ³ /sec.)	2.78	6.50
Current at 1000 m.	0	4.3

TABLE II

Shear Increase Normal to Two Flow Lines
of the Alaskan Stream Due to Latitude Changes

Latitude Change		Shear Increase
from	to	cm./sec. per km.
58° 00' N (B')	54° 25' N (A')	0.50
56° 30' N (B)	53° 45' N (A)	0.40

E. B. Bennett - thesis

Librarian

The figures are marked
'to follow p. -', but I recommend
that they be bound together at the
end — they are at present assembled.

Physics

Figure 1. Geography of the region.

Figure 2a. Station positions.
Figure 2b. Location of sections.

Figure 3. Temperature-depth curve for U of W Station 65.

Figure 4. Depth of surface isothermal layer.

Figure 5. Temperature decrease in thermocline.

Figure 6a. Depth of temperature minimum.
Figure 6b. Temperature minimum values.

Figure 7a. Depth of temperature maximum.
Figure 7b. Temperature maximum values.

Figure 8. Difference between maximum and minimum temperatures.

Figure 9a. Temperature - Section 161.
Figure 9b. Temperature - Section 147.

Figure 9c. Temperature - Section 135.
Figure 9d. Temperature - Section 54.

Figure 9e. Temperature - Section 51.

Figure 9f. Temperature - Section 47.

Figure 10a. Temperature at 10 meters depth.
Figure 10b. Temperature at 50 meters depth.

Figure 10c. Temperature at 100 meters depth.
Figure 10d. Temperature at 200 meters depth.

Figure 10e. Temperature at 400 meters depth.
Figure 10f. Temperature at 600 meters depth.

Figure 10g. Temperature at 1000 meters depth.
 Figure 10h. Temperature at 2000 meters depth.

Figure 11. Salinity-depth curve for POG Station 10.

Figure 12. Salinity at top of halocline.

Figure 13. Depth of the 33.8 ‰ isohaline.

Figure 14a. Salinity - Section 161.
 Figure 14b. Salinity - Section 147.

Figure 14c. Salinity - Section 135.
Figure 14d. Salinity - Section 54.

Figure 14e. Salinity - Section 51.

Figure 14f. Salinity - Section 47.

Figure 15a. Salinity at 10 meters depth.
Figure 15b. Salinity at 50 meters depth.

Figure 15c. Salinity at 100 meters depth.
Figure 15d. Salinity at 200 meters depth.

Figure 15e. Salinity at 400 meters depth.
Figure 15f. Salinity at 600 meters depth.

Figure 15g. Salinity at 1000 meters depth.
 Figure 15h. Salinity at 2000 meters depth.

Figure 16. Density anomaly (σ_t) - Depth curve for POG Station 30.

Figure 17a. Density anomaly (σ_t) - Section 161.
 Figure 17b. Density anomaly (σ_t) - Section 147.

Figure 17c. Density anomaly (σ_T) - Section 135.
 Figure 17d. Density anomaly (σ_T) - Section 54.

Figure 17e. Density anomaly (σ_t) - Section 51.

Figure 17f. Density anomaly (σ_T) - Section 47.

Figure 18a. Density anomaly (σ_T) at 10 meters depth.
 Figure 18b. Density anomaly (σ_T) at 50 meters depth.

Figure 18c. Density anomaly (σ_t) at 100 meters depth.
 Figure 18d. Density anomaly (σ_t) at 200 meters depth.

Figure 18e. Density anomaly (σ_T) at 400 meters depth.
 Figure 18f. Density anomaly (σ_T) at 600 meters depth.

Figure 18g. Density anomaly (σ_T) at 1000 meters depth.

Figure 19. Topography of the 4° C surface.

Figure 20. Topography of the 34 ‰ isohaline surface.

Figure 21. Topography of the isopycnal surface on which σ_t is 27.25.

Figure 22. Geopotential topography referred to 1000 decibars.

Figure 23. Location of stations used for dynamic calculations.

Graphical Representation of D.N.M. Extensions

Figure 24. Graphic methods for D.N.M. extension.

Figure 25. Specific volume anomaly (δ) versus logarithm of depth D.N.M. extension.

Figure 26. Approximate regional occurrence of velocity profile types and profile types.

Figure 27. Surface winds roses, August.

Figure 28. Vector resultant wind field.

Figure 29. Specific volume anomaly (δ) versus logarithm of depth for U of W Stations 30 and 31.

Figure 30. Specific volume anomaly (δ) at 2000 meters depth.

Figure 31. Specific volume anomaly (δ) - Section I, Section II.

Figure 32. Volume transport, 0 - 2000 decibars.

Figure 33a. Volume transport, 0 - 200 decibars, D.N.M. 200 db.
Figure 33b. Volume transport, 0 - 200 db., D.N.M. 2000 db.

Figure 34. Geopotential topography, August 1950.

Figure 35. Surface current across Alaskan Stream due to shear.

Figure 36. Bottom profile between Unimak Island and U of W Station 52; and distribution with depth of specific volume anomaly (δ) north and south of U of W Station 52.

Figure 37. Bottom topography and seamount chains.

Current Deviation by Submarine Ridge

Figure 38. Deflection of current by a submarine ridge.

Figure 39. The 2000 and 3000 meters depth contours of the seamount chains.