THE ALASKAN STREAM

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ABSTRACT

The general oceanographic features and continuity of the Alaskan Stream are discussed using data obtained during May through August 1959. The Alaskan Stream is defined as the extension of the Alaska Current which flows westward along the south side of the Aleutian Islands. It is continuous as far westward as long. 170°E where it divides sending one branch northward into the Bering Sea and one southwestward to rejoin the eastward flowing Subarctic Current.

Observed westward velocities near Atka and Adak Islands were in excess of 100 cm/sec, but maximum geostrophic velocities (referred to 1000-m level) of only 30 cm/sec were obtained from station data. Volume transport, computed from geostrophic currents, was approximately 6×10^6 m^3/sec.

Evidence is presented that the Alaskan Stream is driven primarily by the action of wind stress. The observed narrowness of the stream and continuity of transport also support the view that it is a western boundary current related to the general distribution of wind stress.

CONTENTS

	Page
ACKNOWLEDGMENTS	1
INTRODUCTION	1
Background	2
Previous studies	2
Sources of data	3
BATHYMETRY	5
Island chain	5
Aleutian Trench	5
DISTRIBUTION OF PROPERTIES	5
Salinity	5
Vertical distribution	5
Surface distribution	5
Temperature	7
Vertical distribution	7
Temperature-minimum stratum	8
Temperature-maximum stratum	8
Temperature distribution on sigma-t surface 26.90	9
CURRENTS	10
Direct current data	10
Drift of MV *Pioneer*	10
U.S. Coast and Geodetic Survey drogue studies	11
Japanese drift floats	11
Relative currents	11
Dynamic topography, 0/300 m	11
Dynamic topography, 300/1,000 m	12
Dynamic topography, 0/1,000 m	12
TRANSPORT	12
Relative transport	12
Transport, 0 to 1,000 m	14
Wind-driven transport	14
Sea level pressure and wind stress	15
Ekman transport	17
Total transport	17
Comparison of theoretical and relative transports	18
CONCLUSIONS	18
LITERATURE CITED	19

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INTRODUCTION

The Alaska Peninsula and Aleutian-Komandorski island chain form a partial land barrier across the northern North Pacific Ocean that has a marked effect on the oceanography of the Subarctic Pacific region. A dominant oceanographic feature is the large westward transport of water along the southern side of the chain. Twenty years ago this westward flow was relatively unknown and not considered of sufficient magnitude to be included as one of the current systems of the Pacific Ocean (Sverdrup *et al.*, 1942). Even today an accurate description of this flow cannot be found in either recent general texts of the ocean, such as King (1963), or in recent comprehensive reviews of the oceanography of this region—Fleming (1955), Muromtsev (1958), Uda (1963), and Dodimead *et al.* (1963).

The purpose of this study is to describe the nature and extent of this westward flowing current, the Alaskan Stream, and to evaluate its importance as a major current system of the Pacific Ocean.

Background

According to Sverdrup *et al.* (1942), water between the Aleutian Islands and lat. 42°N is formed by mixing

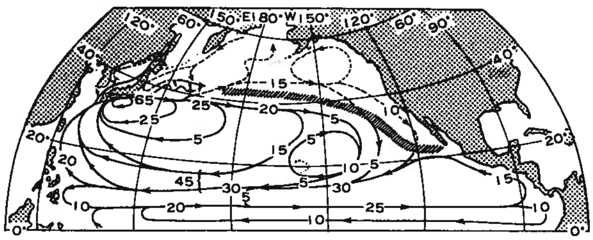

FIGURE 1. Transport chart of the North Pacific. Numbers indicate volumes transported ($m^3/sec \times 10^6$) above 1500 m depth. Dashed lines show cold currents: solid lines show warm currents. (From Sverdrup *et al.*, 1942.)

of the Kuroshio and Oyashio Current systems, and flows eastward as the Aleutian (Subarctic) Current[1]. Near long. 180°, the Subarctic Current divides. One branch flows northward into the Bering Sea. The main current continues eastward to the North American coast where it again divides, sending one branch southward along the west coast of the United States and the other northward into the Gulf of Alaska where it forms a counterclockwise gyre (Fig. 1).

Recent investigations indicate that the northward branch, which is usually considered as the Alaska Current, divides again near long. 160°W; one part flows southward to rejoin the Subarctic Current and completes the Alaska gyre, while the other continues westward along the Aleutian Islands. The latter current is considered in this paper to be the Alaskan Stream.

The term Alaskan Stream has been used before, but has not been well defined. Using data obtained near Unimak Pass (long. 165°W) from several oceanographic stations in 1935 and surface observations in 1938, Goodman and Thompson (1940) defined the Alaskan Stream as a water mass having relatively permanent boundaries and characterized by dilute water believed to be carried into the area as a stream or eddy from the Gulf of Alaska. They attributed the term Alaskan Stream to Schott (1935), but Schott (plate XXX) gives the name as Alaska *Ström* which also translates as Alaska Current, and apparently refers to the Alaska Current rather than to a separate system.

Since these early investigations, this stream or water mass has been denoted in various ways. Doe (1955) identified water masses in the eastern Pacific Ocean and delineated a Coastal Water Mass southward of the Alaska Peninsula; Fleming (1955) divided the area under consideration into two regions, Alaska Coastal and Western Gyral; Kitano (1958) referred to the current south of the Aleutian Islands as a Returning Flow; and Muromtsev (1958) denoted the current moving northward around the Gulf of Alaska and into the Bering Sea through passes in the eastern Aleutian Islands as the Alaska Current.

Bennett (1959) revived use of the term Alaskan Stream. Using data collected by various agencies during 1955, he defined it as that stream south of the Alaska Peninsula, adjacent to the coast and about 100 miles wide. However, the 1955 data were not adequate to define the extent of the Alaskan Stream west of long. 160°W.

Previous Studies

Although many oceanographic cruises had been conducted prior to 1955 in the general vicinity of the Aleutian Islands, most were exploratory. Many consisted of a single line of stations through the area (e.g., Japan Agricultural Technical Association, 1954). Some results of these early cruises have been summarized by Fleming (1955) and by Dodimead *et al.* (1963).

Northward flow through Tanaga Pass was found by Barnes (1936) and further discussed by Barnes and Thompson (1938). Ratmanoff (1937) noted a northward flow through most Aleutian passes, and indicated that the main flow was northward between Attu Island and the Komandorski Islands.

Mishima and Nishizawa (1955) used the limited data from the 1953 and 1954 cruises of the *Oshoro Maru* (which were supplemented with data from the 1929 cruise of the *Carnegie*, the 1933 cruise of the *Gannet* and the 1935 cruise of the *Itsukushima Maru*) to describe conditions in the Aleutian area. They reported that a warm water mass of low salinity flowed westward south of the Aleutian Islands and extended to long. 165°E and that there was a flow northward into the Bering Sea through several shallow channels between the Aleutian Islands, but that the main current turned north or northeastward into the Bering Sea after passing over the Aleutian ridge near Attu Island. They also noted that a large clockwise eddy of this warm water was formed to the south or southwest of Attu Island. This analysis was a reasonably good description of the western terminus of the Alaskan Stream, but because of the few stations and the long time span more extensive observations were needed to document these conclusions.

Solutions to problems concerning the distribution and migration routes of the Pacific salmon which con-

[1] The term Subarctic Current is used in this paper.

fronted the International North Pacific Fisheries Commission in 1953 required knowledge of current systems and the distributions of water properties in the northern North Pacific Ocean. Large catches of salmon near the Aleutian Islands focused attention on that area. Kitano (1958) referred to the current south of of the Aleutian Islands as a Returning Flow with a strong northward component at long. 175°W and 180°, and stated that westward of long. 180° the warm water may occur intermittently as a Cut-Off Warm Water Mass. However, Hirano (1961) reported that water having a relatively high temperature and low salinity flows westward along the southern coast of the Aleutian chain, part going northward and entering the Bering Sea, mainly through the area between Attu Island and the Komandorski Islands, and that a major portion of this water proceeds westward until it turns southwestward off the Kamchatka Peninsula.

Uda (1963), using data available for the years 1955 to 1958, defined an Alaska Stream Extension that developed westward of Unimak Island in spring and summer and extended out to the end of the Aleutian chain by August. He also mentions that water in the Extension which moves northward through the Aleutian passes is affected by a South Aleutian Warm Current; the origin and extent of this current is not clear.

During spring and summer 1957, extensive temperature observations were obtained southward of the Aleutian Islands (Favorite and Pedersen, 1959a) which showed continuity of a warm (greater than 4.0°C) water mass from long. 160°W to 175°E (Callaway, 1963). During 1958, observations of temperature, salinity and dissolved oxygen were obtained (Favorite and Pedersen, 1959b) which showed the distribution of these properties from long. 155°W to 175°E (Dodimead and Favorite, 1961), but the station spacing was not adequate to show continuity of water transport. Based upon information obtained from these cruises, plans were made to obtain more extensive and adequate data in 1959.

Sources of Data

Between May and August of 1959, oceanographic observations were obtained aboard two chartered fishing vessels: MV *Pioneer* and MV *Tordenskjold* (Favorite *et al.*, 1961). Oceanographic work was limited to observations that did not interfere with the fishing operations. Samples were taken at standard intervals from the surface to either 300 or 1,000 m at 30-mile intervals southward of the Alaska Peninsula and Aleutian Islands along seven meridians: 160°W, 165°W, 170°W, 175°W, 180°, 175°E, and 171°E (Fig. 2).

Chemical analyses were performed at a temporary laboratory established at the U.S. Naval Air Station, Adak, Alaska, and at the Bureau of Commercial Fisheries Biological Laboratory, Seattle, Washington. Calculation of the anomalies of dynamic height was accomplished by machine computation.

This paper is based primarily upon these data. Additional data from 1959 cruises of the *Hokusei Maru* and *Oshoro Maru* (Faculty of Fisheries, Hokkaido University, 1960) have been utilized to describe conditions west of long. 172°E, and data from the 1959 cruise of the HMCS *Oshawa* (Fisheries Research Board of Canada, 1959) have been used to determine transports south of Kodiak Island, long. 150°W.

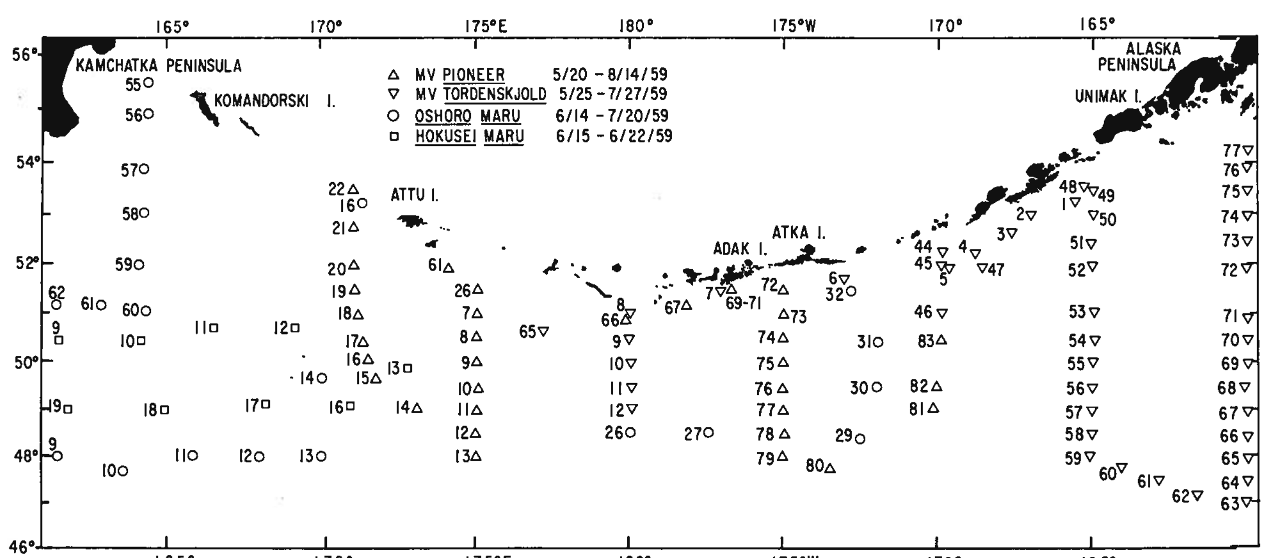

Figure 2. Location of oceanographic stations, 1959.

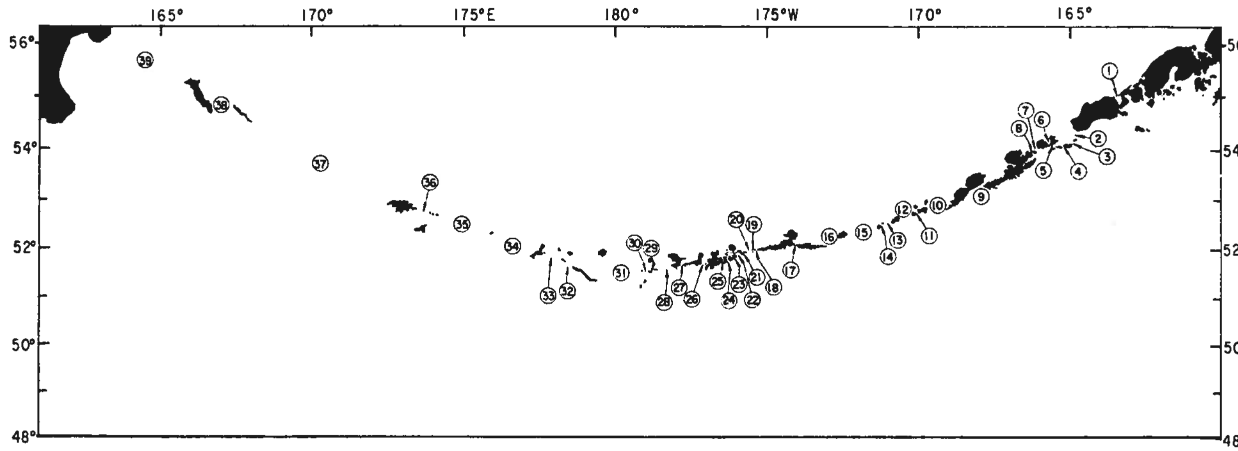

FIGURE 3. Aleutian-Komandorski island chain showing major passes (see Table 1).

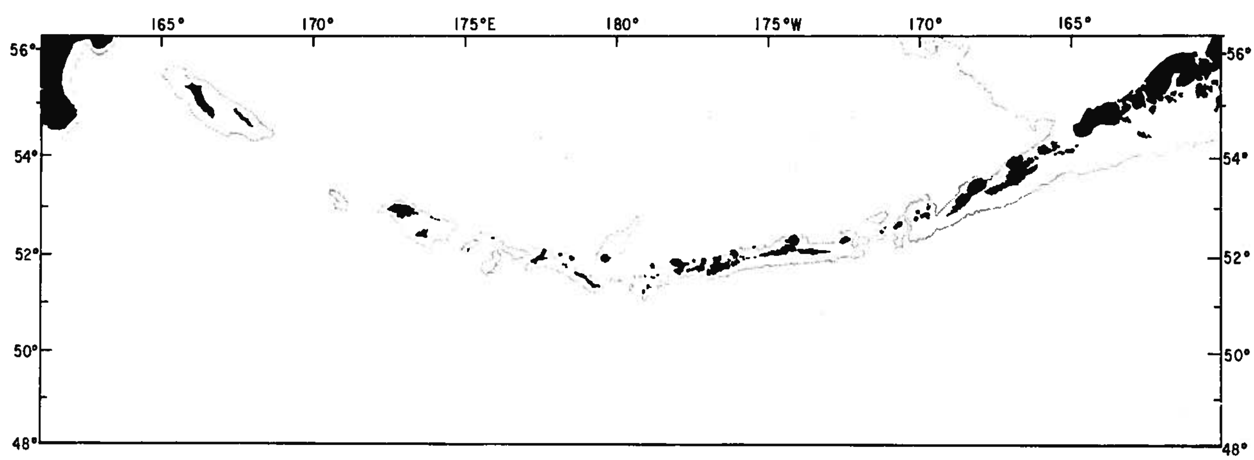

FIGURE 4. Aleutian-Komandorski island chain showing configuration of 300-m depth contour (see Table 1).

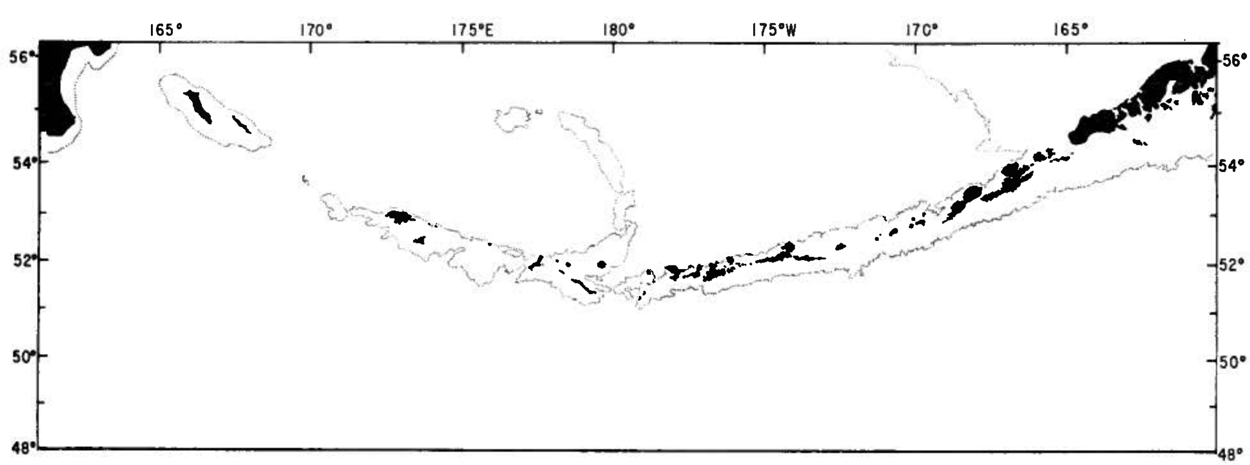

FIGURE 5. Aleutian-Komandorski island chain showing configuration of 1000-m depth contour (see Table 1).

BATHYMETRY

Island Chain

The Aleutian-Komandorski island chain, which extends from the Alaska Peninsula to the Kamchatka Peninsula, separates the North Pacific Ocean from the Bering Sea. Exchange of water can take place in 39 of the major passes (Fig. 3). Only six passes—Yunaska (13), Amukta (15), Amchitka (31), Buldir (35), Near (37) and Kamchatka (39)—have sill depths exceeding 300 m (Fig. 4), and only Amchitka, Near and Kamchatka passes have sill depths exceeding 1,000 m (Fig. 5). Only the passes between the Alaska Peninsula and Attu Island are considered in this paper. Cross sectional areas and sill depths of these passes are tabulated (Table 1). Areas have been obtained by planimetering vertical sections based on bathymetric data obtained from various U.S. Hydrographic Office charts. The total cross sectional area eastward of Attu Island, 130 km^2, is in good agreement with that obtained by Udintsev *et al.* (1959).

Table 1. Cross-sectional areas and sill depths of major passes or straits through the Aleutian Islands.

Location	Pass or Strait[1]	Area ($m^2 \times 10^6$)	Max. sill depth (m)
163°W–165°W			
Alaska Peninsula to Krenitzin Islands	1 False	0.1	35
	2 Unimak	0.9	60
	3 Ugamak	0.2	45
165°W–170°W			
Tigalda Island to Herbert Island	4 Derbin	0.1	50
	5 Avatanak	0.4	80
	6 Akun	0.1	10
	7 Akutan	0.2	70
	8 Unalga	0.1	45
	9 Umnak	0.2	50
	10 Samalga	3.9	200
	11 Chuginadak	1.0	210
170°W–175°W			
Herbert Island to Atka Island	12 (Herbert)	4.8	275
	13 (Yunaska)	6.6	457
	14 Chagulak	0.3	65
	15 Amukta	19.3	430
	16 Seguam	2.1	165
	17 Amlia	0.1	25
175°W–180°			
Atka Island to Amchitka Island	18 Atka	0.2	35
	19 (Oglodak)	0.1	5
	20 Fenimore	0.2	35
	21 Tagalak	0.1	35
	22 Chugul	0.6	85
	23 Umak	0.1	20
	24 Little Tanaga	0.1	30
	25 Kagalaska	0.1	25
	26 Adak	0.5	60
	27 Kanaga	0.1	30
	28 Tanaga	3.6	235
	29 (Ogliuga)	0.1	10
	30 (Kavalga)	0.3	55
	31 Amchitka	45.7	1155
180°–173°E			
Amchitka Island to Attu Island	32 Oglala	0.8	25
	33 (Rat)	0.6	15
	34 (Kiska)	6.8	110
	35 Buldir	28.0	640
	36 Semichi	1.7	105
173°E–163°E			
Attu Island to Kamchatka Peninsula	37 (Near)	239	2000[2]
	38 Komandorski	3.5	105[2]
	39 Kamchatka	335.3	4420[2]

[1] Numbers given correspond to those shown in Figure 3. Parentheses indicate that no geographic name has been assigned.
[2] From Udintsev *et al.* (1959). Maximum depth given.

Aleutian Trench

The Aleutian Trench occurs about 170 km south of the Aleutian-Komandorski island chain and is characterized by depths which vary from 5,000 to over 7,000 m. It extends from the Gulf of Alaska to the Komandorski Ridge, near long. 166°E.

The Komandorski Ridge, indicated by the 4,000 m contour, extends over 700 km southward of the Komandorski Islands (Fig. 6). Two prominent features are the Detroit and Klamath Table Mounts, which rise to within 2,814 and 2,770 m, respectively, of the sea surface.

Eastward of long. 175°E, the bottom topography southward of the Aleutian Trench is relatively uniform, with depths of approximately 5,000 to 6,000 m.

DISTRIBUTION OF PROPERTIES

Salinity

Vertical Distribution

Northward of lat. 42°N, an excess of precipitation over evaporation and an extensive runoff from coastal areas result in a surface layer of low salinity. Salinity values increase continuously with depth, and a relatively sharp halocline usually occurs between 75 and 150 m. A striking feature of the salinity structure southward of the Aleutian Islands is the ridging of the isohalines (Fig. 7). This ridging is continuous from the Gulf of Alaska westward beyond long. 171°E.

Surface Distribution

The distribution of surface salinity in the region between the Bering Sea and lat. 46°N shows a pronounced salinity minimum south of the Aleutian Islands, which is caused by the westward advection of dilute coastal water from the Gulf of Alaska by the Alaskan Stream (Fig. 8). Between long. 175°W and

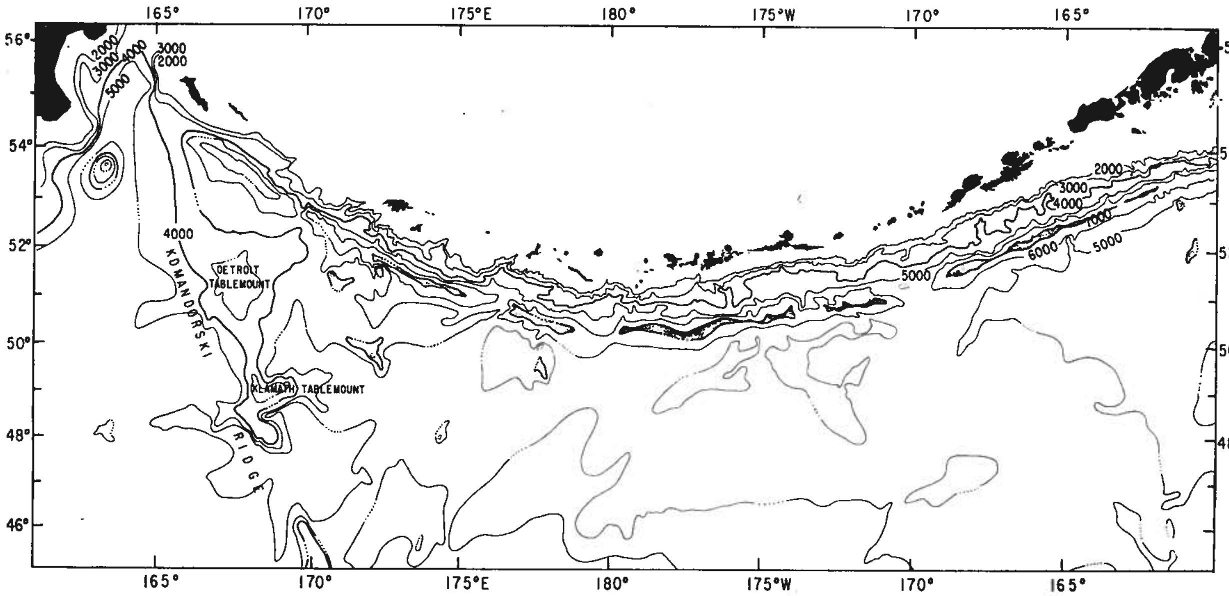

FIGURE 6. Bathymetry southward of Aleutian-Komandorski island chain (depths in meters).

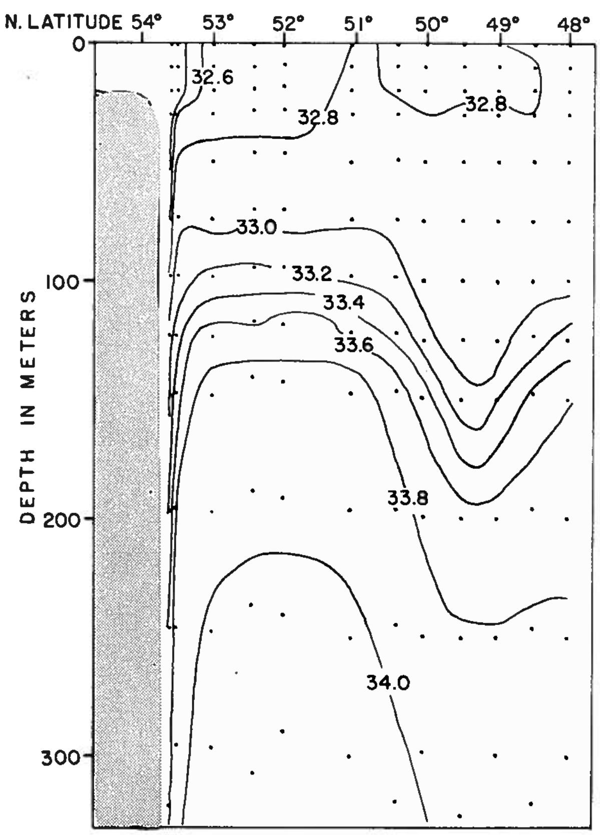

FIGURE 7. Vertical section of salinity (‰) structure at long. 165°W, July 1959 (land shown by stippling).

175°E, the axis of the salinity minimum occurs 50 to 150 km offshore and indicates the absence of any appreciable northward flow through the passes between the Aleutian Islands.

Previous data which are most difficult to reconcile with confinement of the Alaskan Stream southward of the Aleutian Islands were presented by Barnes (1936) and were based upon observations obtained during the cruise of the USS *Gannet* during the period June to August 1933. A significant northward geostrophic flow of approximately 0.3 knot, referred to the 1,000-decibar surface, was reported south of Tanaga Island and the water was shown to flow through Tanaga Pass into Bering Sea. However, the surface salinity in this area was in excess of 33.0‰, which indicates the origin of the water to be the Bering Sea, and the dilute water characteristic of the Alaskan Stream was actually confined well offshore (Fig. 9). A similar surface salinity distribution occurred during 1959 (Fig. 9). Neither the 1933 nor the 1959 salinity data south of Adak Island provide an indication of the abruptness of the front between the Alaskan Stream and water of Bering Sea origin. However, surface observations in this area obtained during August 1963 show a similar surface salinity distribution and a differential of 0.78‰ (33.08 to 32.30) within a 3-mile interval approximately 30 miles south of Adak (Table 2). This indicates that when adequate data are available, the northern boundary of low salinity water characteristic of the Alaskan Stream can be clearly defined, and its location may be found well southward of the Aleutian Islands.

The presence of surface salinities in excess of 33.0‰ south of the Aleutian Islands implies a southward component of surface flow from the Bering Sea through one or more passes, which may or may not flow north-

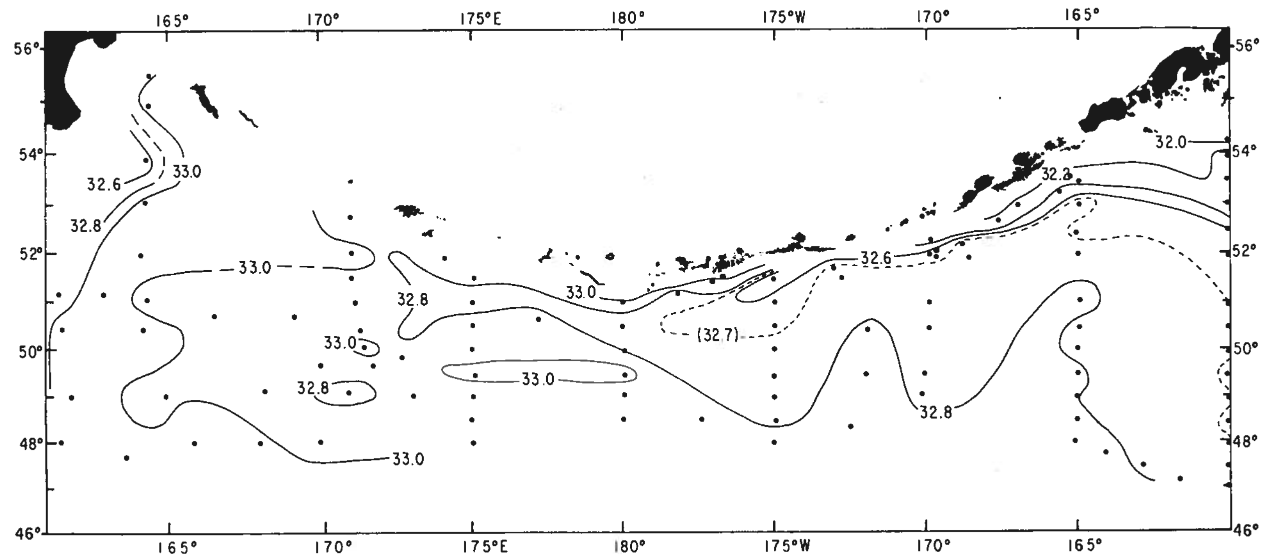

FIGURE 8. Distribution of surface salinity (‰), summer 1959.

ward again through an adjacent pass.

TEMPERATURE

Vertical Distribution

Although the vertical salinity distribution in the central Pacific Ocean north of lat. 42°N exhibits no inversion, the temperature structure is characterized

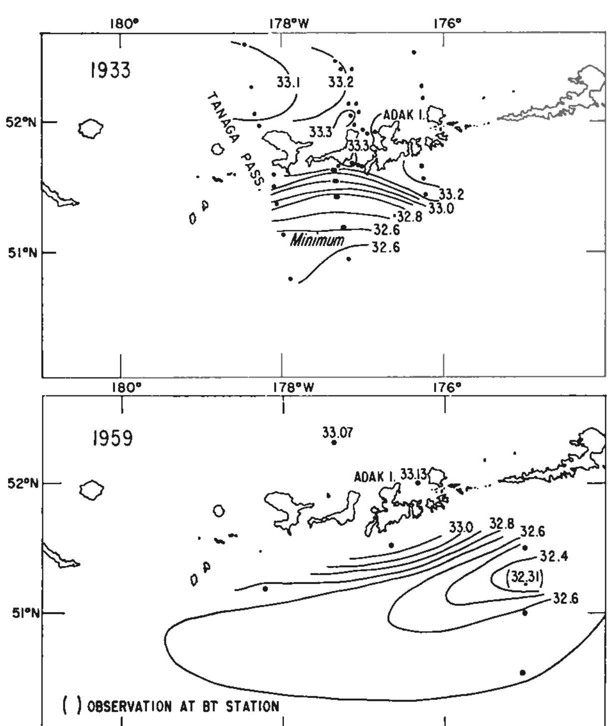

FIGURE 9. Surface salinity (‰) near Adak Island, 1933 and 1959.

TABLE 2. Surface salinity (‰) south of Adak Island, August 1963.

Data	Time	Lat. N	Long. W	Salinity (‰)
8/10	1745	51°42′	176°24′	33.04
8/10	1830	51°36′	176°22′	33.04
8/11	1800	51°15′	176°33′	33.08
8/11	1830	51°12′	176°26′	32.30
8/12	1600	51°02′	176°17′	32.28
8/12	1630	50°57′	176°17′	32.31
8/12	1700	50°52′	176°15′	32.34
8/12	1800	50°44′	176°17′	32.34
8/12	1830	50°49′	176°19′	32.34
8/12	1900	50°36′	176°18′	32.36

by both a minimum and maximum stratum. The temperature-minimum stratum is characterized by two cold cores. One exists at 75 to 125 m and is the result of local seasonal overturn and subsequent heating of the surface layer; it is usually located in the shear zone between the Alaskan Stream and Subarctic Current. The other occurs at approximately 200 to 300 m, and is caused by seasonal overturn in Bering and Okhotsk Seas where the presence of ice permits vertical mixing to a greater depth. This cold water (0°C) is advected eastward in the Subarctic Current, and by the time it reaches the central Pacific Ocean there may be little difference in temperature between it and the shallower cold core caused by local conditions.

The temperature-maximum stratum occurs immediately below the minimum stratum. Temperatures greater than 4.0°C are found over 200 km southward of the Aleutian Islands at a depth of approximately 150 m (Fig. 10). Plakhotnik (1964) has reported the presence of this warm water stratum in the Gulf of Alaska during winter and summer as a relatively permanent feature.

Continuity of these two features can be shown by the distribution of temperature in the temperature-minimum stratum and by the maximum temperatures below the minimum stratum.

Temperature-Minimum Stratum

Several features are evident in the distribution of minimum temperatures (Fig. 11). Temperatures in the northwestern Pacific are several degrees lower than those in the northeastern Pacific Ocean. There is a sharp discontinuity between long. 160°W and long. 165°W near lat. 53°N, and again at long. 165°E near lat. 51°N. The first discontinuity is believed to be associated with the Gulf of Alaska gyre. The second discontinuity is considered to denote the western terminus of the Alaskan Stream. Between lat. 47°N and lat. 50°N, an eastward intrusion of cold water (lower than 3.4°C) extends across the area considered, marking the approximate location of the northern boundary of the Subarctic Current. Finally, there is a discontinuity of isotherms greater than 4°C along the Aleutian Islands; however, isotherms in the temperature-maximum stratum which exists immediately below the temperature-minimum stratum are continuous, implying continuity of the westward flow along the island chain.

Temperature-Maximum Stratum

The distribution of temperature in the temperature-maximum stratum (Fig. 12), suggests continuity of the warm water flowing out of the Gulf of Alaska to long. 171°E and, except for the discontinuity at this longitude which may be due to inadequate sampling rather than to any physical process, it may be considered to

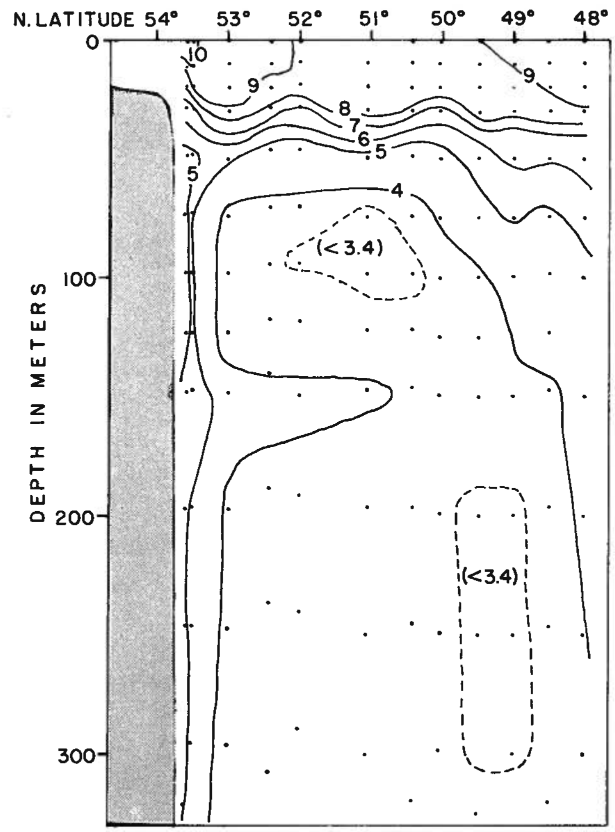

FIGURE 10. Vertical section of temperature (°C) structure at long. 165°W, July 1959 (land shown by stippling).

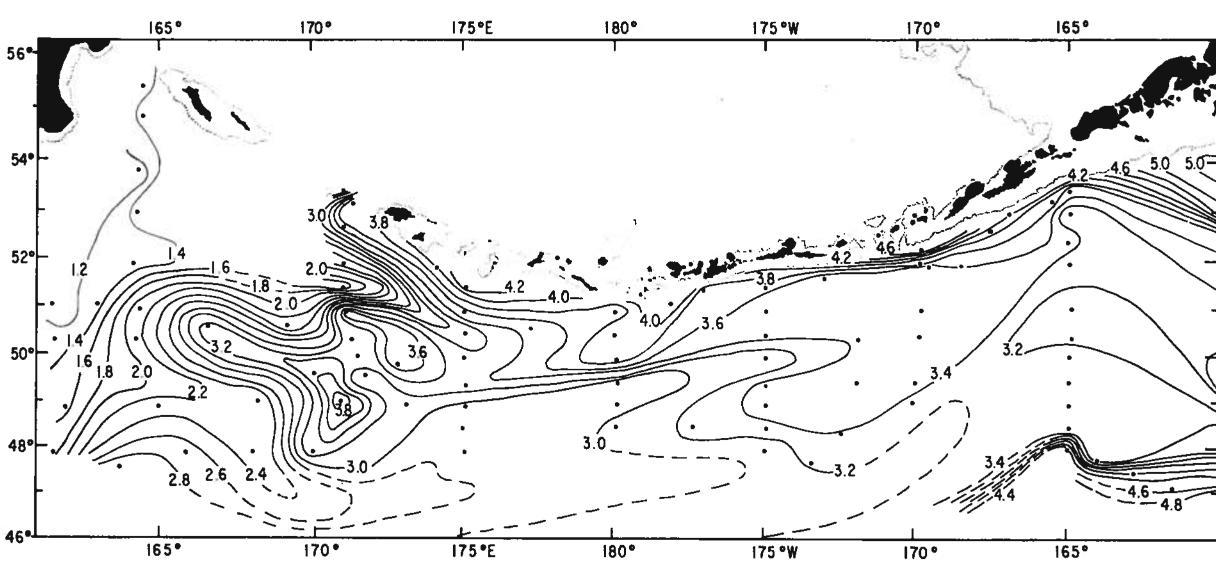

FIGURE 11. Distribution of temperature (°C) in the temperature-minimum stratum, summer 1959.

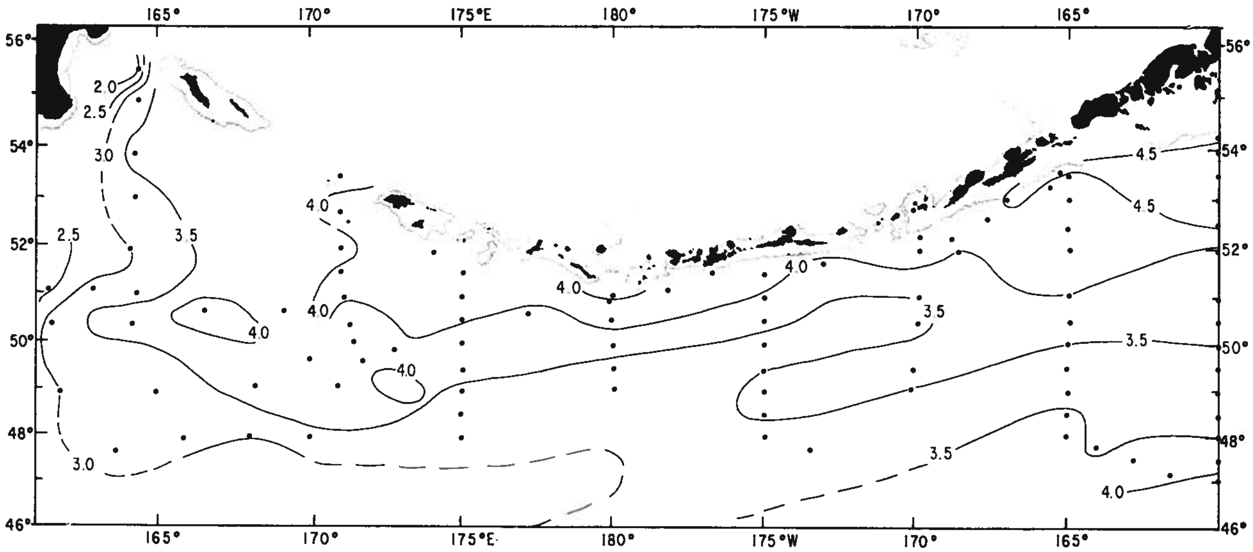

FIGURE 12. Distribution of maximum temperatures (°C) below the temperature-minimum stratum, summer 1959.

extend to approximately long. 162°E. However, the possibility of warm water eddies being pinched off and existing in this area as separate entities cannot be discounted. Also clearly apparent is a tongue of cold water (lower than 3.5°C) south of lat. 51°N which extends from the western Pacific to long. 160°W.

Temperature Distribution on Sigma-t Surface 26.90

To minimize some of the problems involved in interpreting temperature distributions affected by extensive vertical mixing, we can consider the temperature distribution below the depth of seasonal influence. The sigma-t surface 26.90, which is located at a suitable depth to serve our purpose, was selected. In the absence of evidence to the contrary, we may assume that mixing occurs primarily along, rather than normal to, such a surface. The surface selected is an undulating one, usually found between 200 and 400 m throughout the Subarctic region. It commonly lies above 300 m in the Aleutian area, and flow of water of this density could take place through some passes: Samalga (10), Chuginadak (11), Herbert

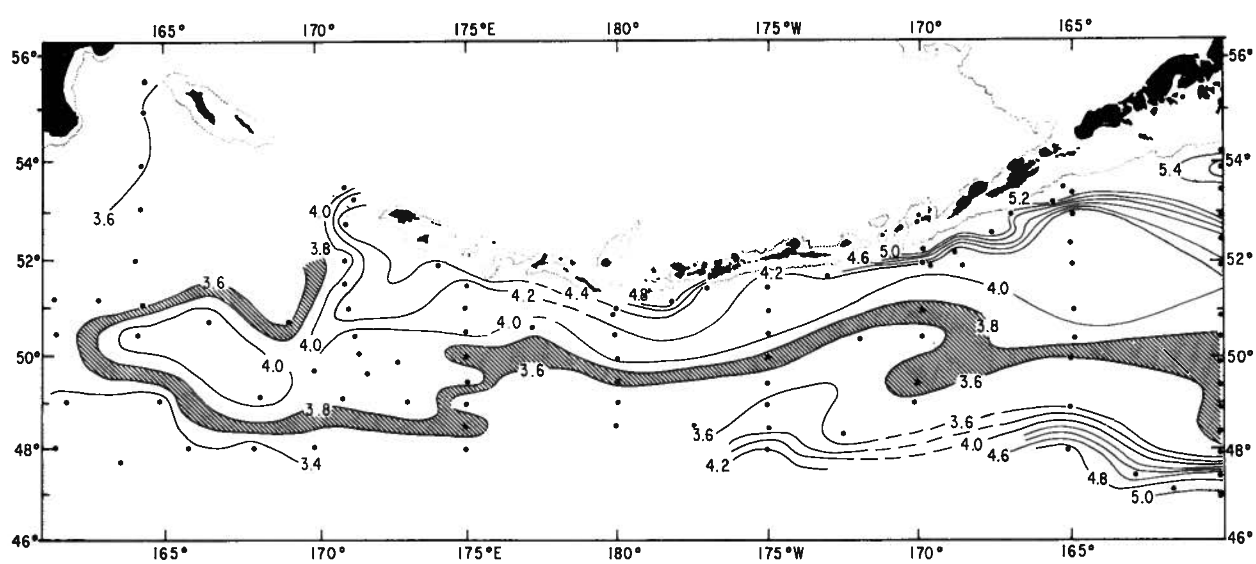

FIGURE 13. Distribution of temperature (°C) on sigma-t surface 26.90, summer 1959. Shaded area shows the approximate southern boundary and western extent of the Alaskan Stream as indicated by temperature distribution.

(12), Yunaska (13), Amukta (15), Tanaga (28), Amchitka (31), Buldir (35), and the remaining western passes, Near (37) and Komandorski (39). However, the 4°C isotherm is continuous from long. 160°W to the westward of long. 170°E (Fig. 13). The approximate location of the boundary between the Subarctic Current and the Alaskan Stream is between the 3.6°C and 3.8°C isotherms.

Analysis of data collected north of the Aleutian Islands during 1959 does not show temperatures in excess of 4°C on the sigma-t surface 26.90. Maximum temperatures are near 3.8°C and are found to extend from long. 171°E to 170°W. It is impossible to ascertain if there is a northward flow through all passes, with a slight lowering of temperature caused by mixing in the passes, or if the main flow is northward only through the western passes, with subsequent eastward advection of this water type along the north side of the Aleutian Islands. Either condition could cause the temperature distribution found on the north side of the island chain.

CURRENTS

The measurement of absolute currents in the ocean is a difficult and important problem. The choice among various proposed theories of ocean circulation may depend upon a comparison of theoretical results with observed ocean currents. As Stommel (1960) says in the concluding sentence of *The Gulf Stream*, " Many of the hypotheses suggested have a peculiar dreamlike quality, and it behooves us to submit them to especial scrutiny and to test them by observation."

Current velocities may be obtained either directly by a metering device or by tracing techniques, or indirectly by analysis of distribution of properties. Direct current observations are difficult to obtain. Fomin (1964) has described some of the obstacles in obtaining direct current measurements. He also presented the qualifications which must be attached to the use of currents calculated by indirect methods. Indirect methods, in general, can provide information only about relative currents, not absolute currents. However, in many parts of the world ocean, indirect methods are the only expedient method of obtaining currents and circulation.

In the Aleutian area the only direct current measurements prior to 1959 appear to have been made in the island passes. The United States Coast Pilot 9—Alaska (U.S. Department of Commerce, 1955) states:

> In the past, numerous reports have been received to the effect that the flood currents flowing into the Bering Sea are very much stronger than the ebb currents. These reports have been largely discounted by observations in a number of passages which in general reveal equally strong ebb currents flowing through the passes from the Bering Sea. It is believed that on account of the large diurnal inequality in the current of this region, mariners have been deceived by the long periods of flood currents that occur near the times of the moon's maximum declination.

Both direct current observations obtained by drogue and vessel drift, and relative current observations obtained by use of the geostrophic approximation, are used to describe the currents and transport in the Alaskan Stream.

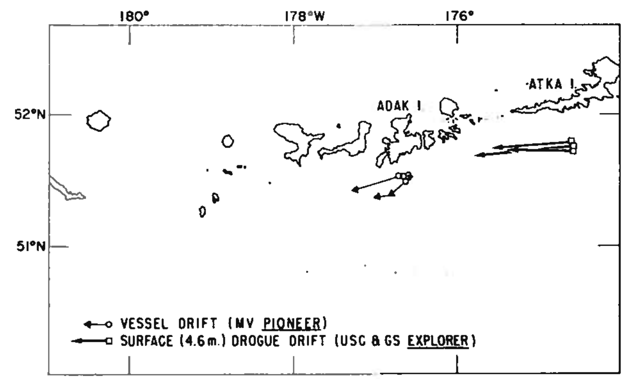

FIGURE 14. Drift of MV *Pioneer* during three consecutive nights (July 26–29, 1959) and drift of parachute drogues (4.6 m depth) released and tracked by USC & GS vessel *Explorer* (June 1959).

DIRECT CURRENT DATA

Drift of MV Pioneer

The set and drift of the MV *Pioneer* were recorded on three consecutive nights at approximately the same time and position south of Adak Island (Fig. 14). During each period, the vessel was moored to about two miles of gillnet. Wind speeds varied from less than 5 m/sec to calm and are not considered to have appreciably affected the vessel's drift, although a small Ekman component may have been present. Maximum drift during an approximate 6-hour period was

TABLE 3. Drift and set of MV *Pioneer* south of Adak Island, July 1959.

Date	Zone time	Lat. N	Long. W	Period (hrs)	Drift (km)	Direction (°T)	Speed (cm/sec)
7/26	1820	51°31′	176′22′				
7/27	0715	51°27′	177°01′	12.9	23	252	50
7/27	1307	51°25′	177°16′	5.9	18	258	85
		Resultant		18.8	41	254	60
7/27	1810	51°31′	176°40′				
7/28	0710	51°29′	176°39′	13	4	163	10
7/28	1815	51°29′	176°38′				
7/29	0705	51°22′	176°49′	12.8	18	226	40
7/29	1215	51°22′	176°58′	5.2	11	270	55
		Resultant		18.0	27	241	40

in excess of 80 cm/sec, and during two periods an average velocity of about 50 cm/sec in a westerly direction occurred (Table 3).

U.S. Coast and Geodetic Survey Drogue Studies

Current measurements by the USC&GS *Explorer* south of Atka Island during June 1959 also indicated a swift westerly flow. Six parachute drogues, three shallow (4.6 m) and three deep (300 m), were released and tracked for over 24 hours. Drogue positions were recorded every two or three hours. The three shallow, or surface, drogues moved at an average velocity of about 75 cm/sec, 270°T (Fig. 14). Maximum speeds in excess of 100 cm/sec were recorded. One of the deep drogues grounded, but the other two moved at an average velocity of over 50 cm/sec, 270°T. The trajectories of all drogues were essentially straight line paths and no evidence of north-south tidal currents was found.

Japanese Drift Floats

Taguchi (1959) conducted an extensive drift float experiment in the western Pacific Ocean during the period in which the oceanographic data under discussion were obtained. Hooks attached to floats facilitated capture in gillnets set each night by the extensive Japanese fishing fleet. The release and recovery points of selected floats (Fig. 15) suggest that the westward surface flow in this area terminates at long. 165°E and that there is a northward surface flow into the Bering Sea west of Attu Island. The presence of the Komandorski Ridge at long. 166°E may affect the flow in this area.

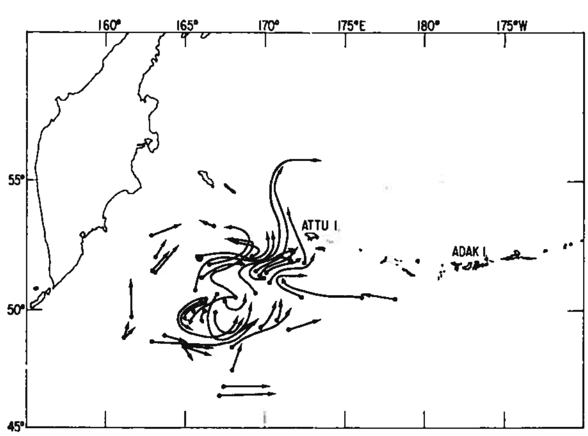

FIGURE 15. Release and recovery points of selected drift floats, May to July 1959 (adapted from Taguchi, 1959).

RELATIVE CURRENTS

The dynamical method permits calculation of relative currents from the distribution of mass, assuming: synoptic data, a balance between the pressure gradient and Coriolis forces, unaccelerated flow, no friction, and a known reference level of no-net-motion. In the absence of *a priori* knowledge of the reference level only velocities relative to the velocity on an arbitrary selected surface are obtained, rather than absolute values.

The dynamic topographies have been calculated from the modified hydrostatic equation:

$$\Delta D = \int_{p_1}^{p_2} \delta \, dp$$

where ΔD=geopotential anomaly, δ=specific volume anomaly, and dp=pressure differential.

The component of relative velocity normal to the line between two stations, A and B, has been computed using the equation:

$$v = \frac{10(\Delta D\text{A} - \Delta D\text{B})}{L\, 2\omega \sin\phi}$$

where v=velocity, ΔD=geopotential anomaly, L= distance between stations, ω=angular velocity of earth (0.729×10^{-4} rad/sec), and ϕ=mean latitude of the stations.

Relative surface currents based on a reference level of 300 m will be considered first, since data from almost all stations extend to this depth. Although these data are not synoptic, they were obtained over a relatively short period of time. The observed values have been used and no attempt has been made to adjust the data to a mean time period, thereby accounting for seasonal heating, as this would not significantly alter the dynamic topography or the conclusions.

Although the distribution of properties makes it appear reasonable that a reference surface of no-net-motion must lie below 300 m, the major speed changes occur in the upper layers and the 0/300 m dynamic topography indicates the dominant features of the circulation.

Dynamic Topography, 0/300 m

As might be expected in a shear zone between two opposing currents, a number of eddies or gyres are suggested by the surface topography, but the limited data, and location and time of sections make it impossible to define the actual extent of each gyre (Fig. 16). Although the 0.440 dynamic meter surface is present from long. 180° to 160°W, it appears discontinuous at several locations. Thus, the 0.460 dynamic meter surface has been used to indicate the northern boundary of the eastward flow south of lat. 51°N, and the southern boundary of the westward flow between lat. 51°N and the Aleutian Islands. The former is the Subarctic Current and the latter is the Alaskan Stream.

There is a general continuity of westward flow in the Alaskan Stream between long. 160°W and 170°E. At long. 170°E the flow turns northward and its effect on the two gyres at long. 171°E and 167°E near lat. 50°N is not readily apparent.

Dynamic Topography, 300/1,000 m

Not all stations extend to 1,000 m, but analysis of the flow at 300 m relative to 1,000 m from the data available also shows the eastward and westward flows described in the previous section, south and north of a clearly defined trough. This indicates that the readjustment of mass extends well below the depth of seasonal influence, which has been shown to be approximately 100 to 150 m (Dodimead *et al.*, 1963). It also implies that the current systems must be relatively permanent.

Dynamic Topography, 0/1,000 m

Because extreme gradients are not found below 300 m, it is believed that data from the coarse grid of stations that extends to 1,000 m adequately define the 300/1,000 m topography, and the 0/300 m topography has been superimposed upon that of 300/1,000 m to give the 0/1,000 m topography (Fig. 17). The continuity of the westward flow in the Alaskan Stream from long. 160°W to long. 170°E is clearly evident. The cyclonic gyre near lat. 50°N, long. 172°E, is more clearly defined, and the validity of the circulation pattern in this area, which indicates a divergence of the Alaskan Stream, is reinforced by the drift float study discussed in the section on direct current data.

The maximum relative speeds, about 30 cm/sec, occurred between *Tordenskjold* Station Nos. 77 and 76 (long. 160°W). Generally, speeds of about 10 cm/sec were calculated for the Alaskan Stream.

TRANSPORT

Geostrophic surface currents relative to 1,000 m have been used to indicate the magnitude of the Alaskan Stream and to mark its boundaries. To describe completely the Alaskan Stream, it is necessary to specify not only the general extent of the current system, but to provide absolute currents and mass transport, and to relate the Alaskan Stream to the general pattern of circulation in the North Pacific Ocean.

Relative Transport

The calculation of relative transport requires a knowledge of the internal distribution of currents in the water column. Relative transport calculated from the observational data by integrating the relative velocities will be considered as a first approximation to the transport in the Alaskan Stream. These calculations are subject to the same qualifications mentioned in the discussion of relative currents, with somewhat more uncertainty than the current speeds, since any error in the reference velocity will be integrated over the water column.

The necessity of occupying station positions according to a predetermined plan introduces another factor which may reduce the apparent observed relative transport below its true value. To obtain the maximum value of the geostrophic transport, data must be obtained at both boundaries of the system,

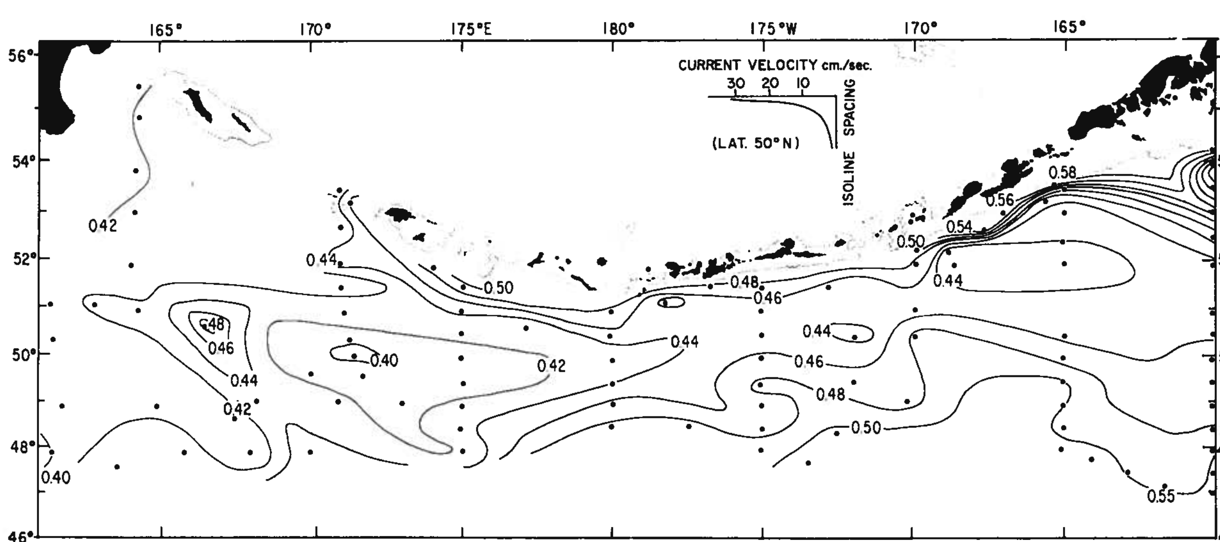

FIGURE 16. Dynamic topography, 0/300 m, summer 1959.

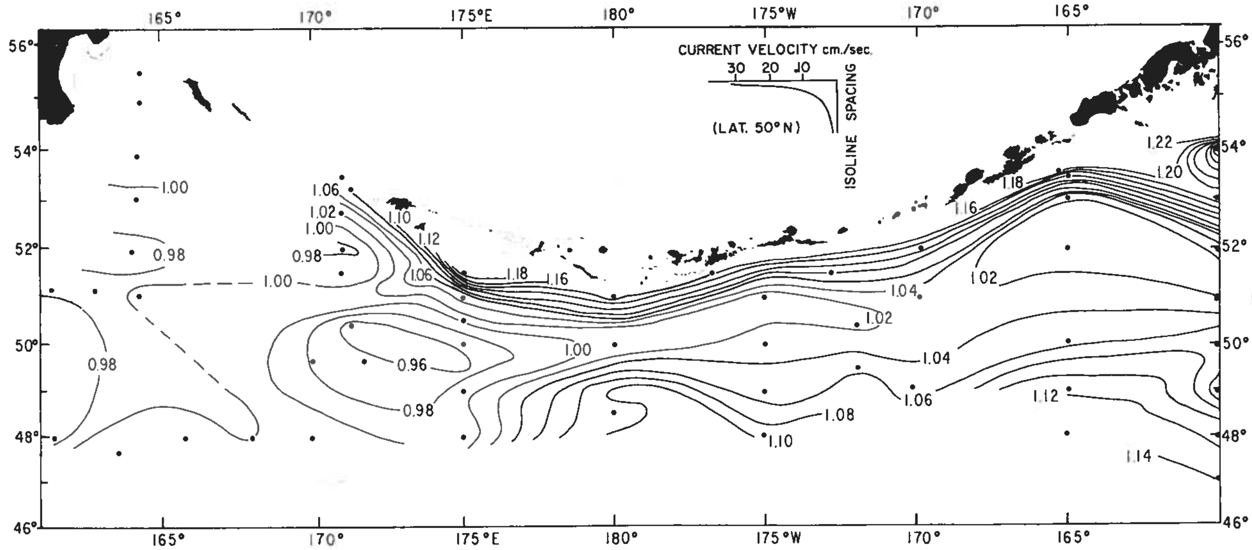

FIGURE 17. Dynamic topography, 0/1000 m, summer 1959.

TABLE 4. Dynamic anomalies at stations used for computations of volume transport[1].

	Long. 160°W		Long. 165°W		Long. 170°W		Long. 175°W	
Depth (m)	Stn. T-77	Stn. T-71	Stn. T-48	Stn. T-52	Stn. T-44	Stn. T-46	Stn. P-69	Stn. P-73
0	0	0	0	0	0	0	0	0
30	.092	.083	.087	.076	.077	.070	.055	.080
50	.146	.130	.137	.118	.122	.112	.090	.124
75	.206	.178	.194	.166	.174	.161	.132	.172
100	.262	.220	.246	.208	.221	.207	.173	.215
150	.365	.289	.339	.276	.302	.287	.253	.286
200	.454	.345	.425	.332	.372	.406	.331	.343
300	.590	.449	.574	.436	(.496)	.458	.479	.447
400	(.699)	.547	(.684)	.532	(.606)	.557	(.602)	.544
500	(.800)	.639	(.784)	.621	(.706)	.650	(.709)	.633
600	(.895)	.725	(.879)	.705	(.801)	.737	(.807)	.717
800	(1.072)	.880	(1.057)	.861	(.979)	.894	(.986)	.871
1000	(1.231)	1.020	(1.216)	1.003	(1.138)	1.035	(1.142)	1.012

	Long. 180°		Long. 175°E		Long. 171°E			
Depth (m.)	Stn. T-8	Stn. T-10	Stn. P-62	Stn. P-10	Stn. O-16	Stn. P-19	Stn. P-18	Stn. P-16
0	0	0	0	0	0	0	0	0
30	.057	.063	.068	.054	.062	.057	.057	.057
50	.092	.102	.105	.090	.100	.092	.095	.094
75	.134	.148	.148	.134	.142	.131	.141	.137
100	.176	.190	.189	.176	.183	.169	.184	.175
150	.257	.261	.269	.245	.262	.238	.260	.239
200	.334	.321	.347	.301	.335	.298	.324	.293
300	.478	.428	.493	.405	.461	.406	.432	.393
400	.601	.526	.617	.500	.571	.505	(.531)	(.486)
500	.708	.618	.727	.589	.672	.597	(.623)	(.573)
600	.806	.705	.829	.672	.762	.684	(.710)	(.656)
800	.985	.865	1.015	.824	.928	.843	(.869)	(.808)
1000	1.141	1.008	1.181	.959	1.084	.986	(1.012)	(.945)

[1] P=*Pioneer*, T=*Tordenskjold*, O=*Oshoro Maru;* see Figure 2.
Parentheses around station data indicate values obtained from adjoining stations which extended to 1000 m, and had similar properties at 200 and 300 m.

the geopotential along the crest and trough of the topography. Otherwise, the section may contain opposing flow and considerably reduced transports would result. Because stations were occupied at predetermined positions, precise values have not been obtained for the entire stream across all sections, but the close station spacing has provided a good continuity of transport across the previously discussed north-south sections. Serial dynamic anomalies are listed for reference (Table 4).

Relative volume transports have been computed using the following equation from Sverdrup *et al.* (1942, p. 463):

$$T = L\int_0^d v dz = 10c \int_0^d (\Delta D_A - \Delta D_B) dz$$

where T=relative volume transport, L=distance between stations A and B, v=geostrophic velocity, dz= depth increment, $c=1/2\,\omega \sin\phi$, and ΔD=geopotential anomaly.

Transport, 0 to 1,000 m

Since geopotentials are essentially streamlines for geostrophic flow, the dynamic topography shown (Fig. 17) indicates the flow pattern which may be expected at the surface in the Alaskan Stream. Since the topography at intermediate depths with respect to 1,000 m is similar to that at the surface, the integrated transport pattern will also be similar to the surface dynamic topography. Stations have been selected which are located approximately along the same geopotentials (1.02 and 1.14 dynamic meters), and geostrophic volume transports have been calculated for the sections between the stations (Fig. 18).

Transport in the Alaska Current, which moves cyclonically around the periphery of the Gulf of Alaska, is based on the 1000-m reference level and is usually 10×10^6 m^3/sec. Calculations based upon data from the HMCS *Oshawa* (Fisheries Research Board of Canada, 1959), show that in 1959 the westward transport in the northern part of the Gulf of Alaska was 9.8×10^6 m^3/sec. The westward transport across long. 165°W of 5.9×10^6 m^3/sec indicates that 3.9×10^6 m^3/sec had returned southward to the Gulf of Alaska gyre eastward of this longitude (a loss of 2.5×10^6 m^3/sec occurs between long. 160° W and 165° W). This southward flow is evident in the dynamic topography (Fig. 17).

The relatively small transport values across the north and south boundaries westward of long. 165°W are probably related more to station position than actual flow. An east-west transport of 6×10^6 m^3/sec occurs between long. 165°W and 175°E. No opposing transport is observed until long. 171°E where an eastward intrusion results in a divergence of flow. Lack of data to 1,000 m and inadequate station locations west of long. 171°E make any presentation of transport based on dynamic topography unreliable beyond this location.

Wind-Driven Transport

The relative transport calculated by the geostrophic approximation provides no information about the cause of the Alaskan Stream, or its relation to the general circulation in the North Pacific Ocean. The assumptions used in reducing the relative transports to absolute values are essentially facts observed about the stream itself, and the true transports still do not

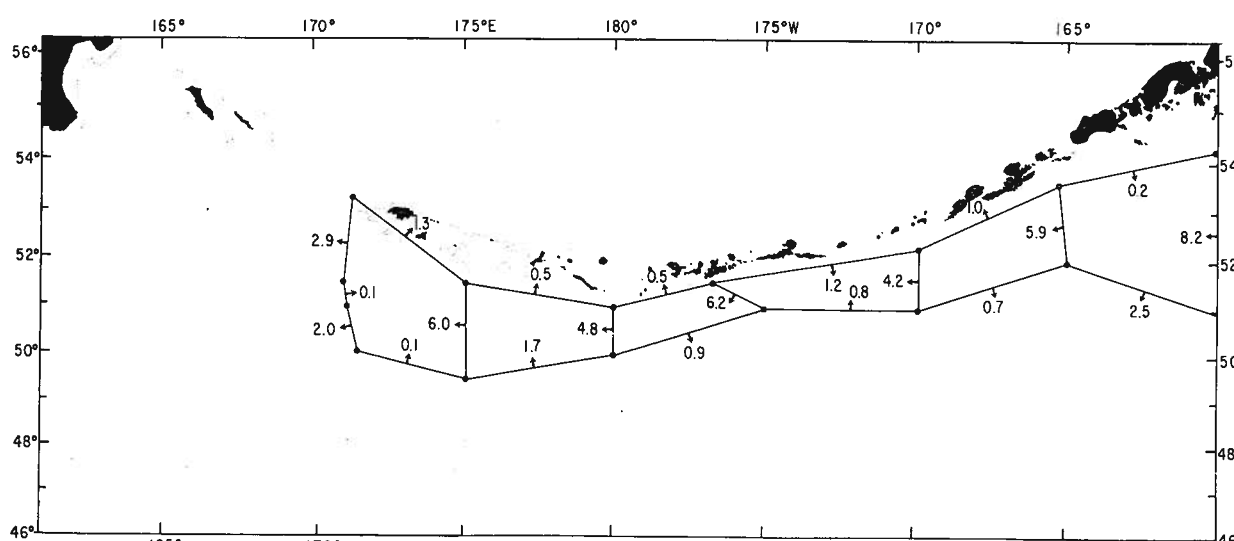

FIGURE 18. Relative transport (m^3/sec$\times10^6$) in the Alaskan Stream, 1959.

provide an indication of the forces that cause the Alaskan Stream.

Some insight into the reasons for the stream appearing as it does can be obtained by consideration of the transport of vorticity within the stream. Since the stream is bounded on both sides by lines of zero velocity, it will contain no relative vorticity and transport no relative vorticity. Separate regions within the stream can, and of course will, possess relative vorticity, but the net relative vorticity across the stream must be zero, regardless of the distribution of vorticity within the stream.

The observed stream moves from higher latitudes to lower latitudes, that is from regions of higher to lower planetary vorticity. Therefore, the total counterclockwise vorticity of the stream is reduced in the direction of the flow. The stream thus serves the function of removing counterclockwise vorticity from the ocean, and the observed features of the stream result from the manner in which the vorticity is removed. An intuitive picture of how this takes place can be presented. As the stream begins to flow southward, the released planetary vorticity appears as an increase in the counterclockwise relative vorticity. Since the stream cannot transport any relative vorticity, this may be compensated for by a production of clockwise relative vorticity produced by a shrinking of the stream. This process will continue until the stream has deepened, narrowed, and increased in velocity such that the friction on the inshore side is great enough to add enough clockwise vorticity to the stream to compensate for the counterclockwise relative vorticity being exchanged from the planetary vorticity. Under these conditions the flow will become steady. Actual values for the distribution of depth, width, and velocity in the stream will depend upon such factors as: the density of the water available for the stream, density of the water through which the stream passes, rate of change of planetary vorticity, the nature of the frictional forces, and the means by which the vorticity is diffused across the stream. The relations which obtain among these factors in the Alaskan Stream are complex and not exactly known. Certain qualitative features may, however, be deduced for the steady stream. As vorticity is introduced at the land boundary, we would expect the greatest shear to take place there, and the cross-stream velocity profile to be asymmetric, with the axis strongly shifted towards the shore. The reference depth will vary from the surface on the interior boundary of the stream to some maximum value near the axis. These features are observed in the Alaskan Stream.

The curl of the wind stress changes sign at about lat. 50°N in the eastern North Pacific Ocean (Fofonoff, 1961), such that for latitudes north of lat. 50°N the effect of the curl of the wind stress is to add counterclockwise relative vorticity to the water. However, since the relative vorticity is negligible, and the water in a major ocean basin is constrained from shrinking by its boundaries, it must turn northward into regions of larger counterclockwise planetary vorticity. If the initial basis for the stream is to provide a physical means of removing vorticity which has been added to the North Pacific Ocean by the wind, then the observed transport in the Alaskan Stream should balance the northward component of the wind-driven circulation in the eastern North Pacific Ocean.

The magnitudes of the transport field of the local wind near the Aleutian Islands, and the wind-driven transport in the eastern North Pacific Ocean were investigated to see if the Alaskan Stream may properly be considered as a return flow for the wind-driven transport. Due to the lack of wind data from the Aleutian region, it was necessary to rely on sea level pressures and calculate the transports from the mean pressure field. The method of Fofonoff (1961) was followed, but extended to obtain more detailed information.

Sea Level Pressure and Wind Stress

The month of July was selected as the mean time period of the oceanographic data; sea level pressures and winds during July 1959 were derived from 6-hourly weather maps furnished by the Seattle Office of the U.S. Weather Bureau. A relationship between the pressure and wind was obtained using observed wind vectors in the vicinity of lat. 45°N to 50°N, long.

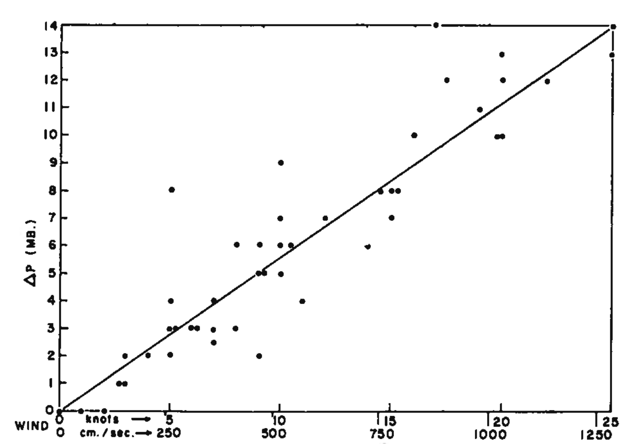

FIGURE 19. Relationship between east-west component of surface winds (rotated 15° to the right of the downwind direction) *versus* north-south sea-level pressure gradient in the vicinity of lat. 45°N–50°N, long. 180°–170°W, July 1959.

180° to 170°W. These vectors were rotated 15° to the right of the surface downwind direction so that they paralleled the isobars, and the resolved zonal components were then plotted against the respective north-south pressure gradient between two grid points. Since most of the weather data are derived from widely spaced and intermittent ship reports, a large scatter was present; however, a general linear relationship was obtained (Fig. 19).

Sea level pressure grid points presented by Fofonoff (1961) were considered too widely spaced to be effective in determining the effect of local wind stress in the narrow region of the Alaskan Stream. Values of sea level pressure at 38 grid points, separated by 5° of latitude and longitude, were interpolated from each 6-hour chart for July 1959. Average weekly, biweekly, and monthly pressure distributions were drawn to determine if any unusual conditions occurred in the Aleutian area that might have been masked. None were apparent, and the detailed chart of mean

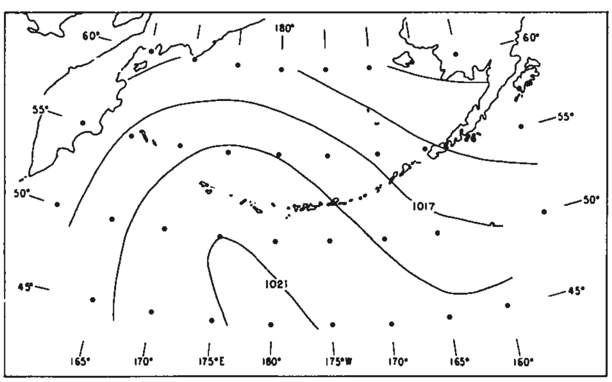

FIGURE 20. Mean sea level pressure during July 1959, determined from the average of pressures shown on 6-hour weather charts at the indicated grid points (contour interval 2 mb).

sea level pressure for July and the locations of the grid points are shown in Figure 20.

The resulting pressure distribution is almost identical to that presented by Fofonoff (1961) for July 1959. The general stress law:

$$\tau = \rho \gamma^2 v^2$$

(where τ=surface wind stress, dynes/cm^2; ρ=density of air, 1.22×10^{-3} gm/cm^3; γ^2=non-dimensional drag coefficient taken as 1.9×10^{-3} for an average wind speed of 13 m/sec (Deacon and Webb, 1962, p. 57); and v=wind speed in cm/sec) has been used to compute wind stress and the monthly sea level pressures have been used to represent winds. Introduction of the mean monthly pressure produces an error when the above stress law is utilized because the monthly root-mean-square of the individual daily mean usually will not be equal to the square of the monthly mean. The magnitude of this error was investigated. Daily 6-hour pressure gradients between meridional grid points were averaged, squared, and summed for the month. The monthly root-mean-square values of pressure gradients between north-south grid points were summed to establish the new pressure field. The resulting root-mean-square pressure field (Fig. 21) shows a considerable increase in the pressure gradients from that computed using mean monthly pressures determined from 6-hourly observations.

The validity of the relationship between the wind speed and pressure gradient can be supported by 6-hourly observations of wind at a shore station. The monthly root-mean-square wind velocity vector at Adak Island (long. 178°W) for July 1959 gives a resultant wind of 8.2 knots from the southwest, the wind

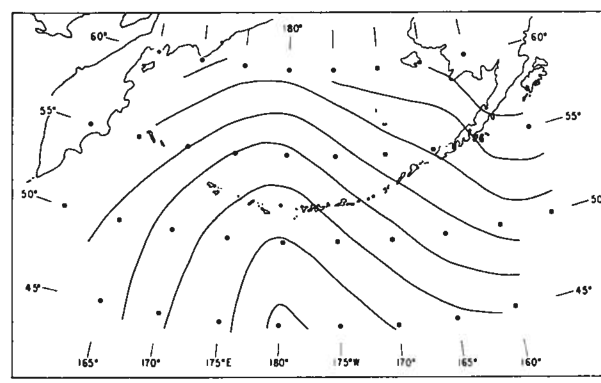

FIGURE 21. Root-mean-square sea level pressure field, July 1959 (contour interval 2 mb).

from the root-mean-square pressure field between lat. 50° and 55°N at this longitude is a 7.0 knot westerly wind. The wind speeds are in good agreement. Lack of exact agreement in wind direction may have

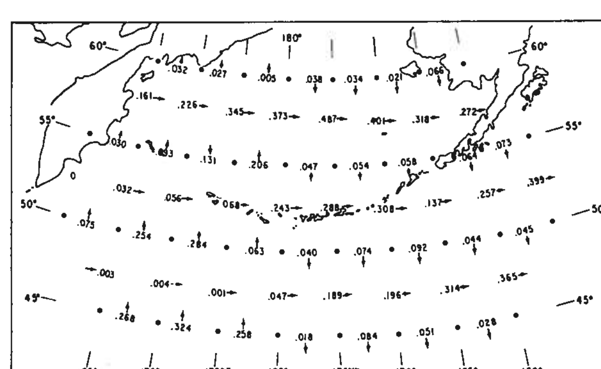

FIGURE 22. Zonal and meridional components of wind stress (dynes/cm^2), July 1959.

been caused by the mountainous topography of Adak Island in the vicinity of the reporting station.

The zonal (τ_x) and meridional (τ_y) wind stresses between grid points were computed by including a factor, K ($K=1.00$ at lat. 47.5°N, 0.93 at lat. 52.5°N, and 0.88 at lat. 57.5°N), in the relationship between wind velocity and pressure gradient ($v\alpha\Delta p$) to compensate for the latitudinal change in Coriolis parameter from the location the relationship was obtained

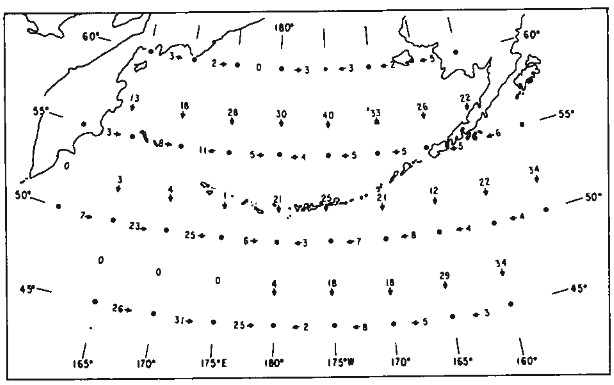

FIGURE 23. Zonal and meridional components of Ekman transport (m^3/sec/km $\times$ 10), July 1959.

(lat. 45°N to 50°N) and by rotating the wind 15° to the left of the downwind direction of the pressure distribution (Fig. 22).

Ekman Transport

The Ekman transport has been derived directly from the wind stress and Coriolis parameter. Zonal and meridional Ekman transports in units 10 m^3/sec/km were derived (Fig. 23) using the relations:

$$U=\frac{\tau_y}{f} \text{ and } V=-\frac{\tau_x}{f}$$

where U, V=zonal and meridional components of Ekman transport, respectively; τ_x, τ_y=zonal and meridional components of wind stress, dynes/cm^2, respectively; and f=Coriolis parameter, rad/sec.

An interesting feature of the results is that all significant meridional components are to the south. In the vicinity of the Aleutian passes, maximum transport occurs between long. 170°W and 180°. Considering the total meridional transport, a southward transport of approximately 0.09×10^6 m^3/sec is obtained between these two longitudes. This is about an order of magnitude less than the transport indicated by the distribution of mass. However, it could be significant in confining the Alaskan Stream southward of the island chain. Another significant result is that, in general, components of Ekman transport in the vicinity of the Alaskan Stream are very small in comparison to the transports of 6×10^6 m^3/sec obtained from the distribution of mass.

Total Transport

Total transport is a function of the curl of the wind stress which may be expressed as:

$$\mathrm{curl}_z\,\tau = \delta\tau_y/\delta_x - \delta\tau_x/\delta_y.$$

The zonal wind stress gradient, $\delta\tau_x/\delta_y$, was found at each grid point by taking the difference between the values of τ_x immediately north and south of the grid points. The meridional wind stress gradient, $\delta\tau_y/\delta_x$, is obtained in a similar manner, in the east-west direction, but this component must be adjusted because of the northward convergence of the meridians. The following factors were used: lat. 45°N—1.409, lat. 50°N—1.551, lat. 55°N—1.739, lat. 60°N—1.997. Subtraction of the two values at each grid point gives the value of $\mathrm{curl}_z\tau$ (Fig. 24).

The southern boundary of the Alaskan Stream has been shown to be a region of low flow. Assuming the

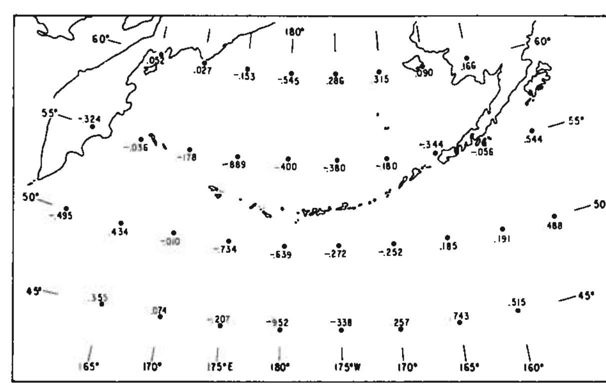

FIGURE 24. Curl of wind stress (dynes/cm^2/km $\times$ 10^{-3}), July 1959.

boundary may be approximated by the latitude circle, 50°N, meridional components of total mass transport may be computed between grid points along this circle using the following relationship:

$$T_m=\mathrm{curl}_z\,\tau/\beta$$

where T_m=total meridional transport of mass across a unit length of a latitude circle, and β=rate of variation of Coriolis parameter along a meridian.

The effects of local wind stress on the Alaskan Stream may be ascertained by considering the island chain as a solid boundary and the transport across long. 155°W to be zero. Then integrated transports resulting from curl of the local wind stress can be obtained by continuity principles (Fig. 25). A westward transport is obtained from long. 155°W to long. 175°W which has a maximum of 1.5×10^6 m^3/sec at long. 165°W. Between long. 170°E and 180° an

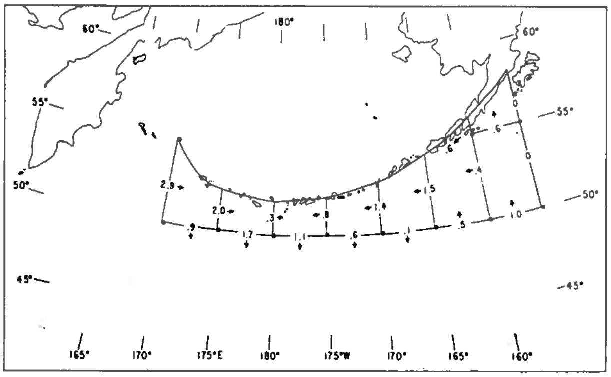

FIGURE 25. Integrated total transports ($m^3/sec \times 10^6$) from local wind stress, July 1959.

eastward transport is present which has a maximum of 2.9×10^6 m^3/sec at long. 170°E.

In order to determine the total integrated transport in the Alaskan Stream, a value for total integrated transport across long. 155°W between lat. 50°N and the Alaska Peninsula is required. This has been computed by Fofonoff (1961, p. 140) and is 5.2×10^{-6} m^3/sec. If this boundary value is used, instead of zero as above, the total integrated transport south of the Aleutian Islands is found to reach a maximum of 6.7×10^6 m^3/sec at long. 165°W and to gradually decrease to 2.3×10^6 m^3/sec at long. 170°E (Fig 26).

Comparison of Theoretical and Relative Transports

The relative transport has been computed from the geostrophic approximation, and thus cannot be compared with the theoretical total transport obtained from the wind stress until the effect of the Ekman transport is considered. If we now define geostrophic transport as that part of the wind-driven transport which will appear associated with a distribution of mass, the meridional component of the geostrophic transport across a latitude circle is equal to the meridional component of total transport minus the meridional component of Ekman transport. Since the maximum Ekman transport between any of the 5° longitudinal segments indicated along lat. 50°N has been calculated to be 0.09×10^6 m^3/sec (based on Fig. 23), the direction of meridional geostrophic transport across any of these segments is the same as that indicated by the total transport, and the values are not significantly changed. Thus, the zonal components of geostrophic transport may be considered equivalent to the zonal components of total transport and may be compared directly to the relative transport calculated from the observed distribution of mass.

West of long. 180° the zonal geostrophic transport induced by local wind stress opposed the westward flowing Alaskan Stream (Fig. 25). Reasonably good agreement between these two transports is evident by comparison of Figures 18 and 26.

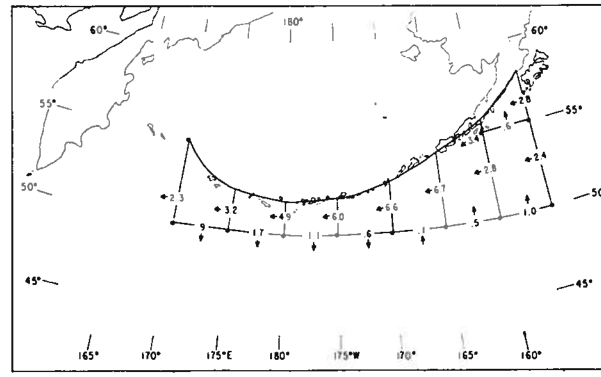

FIGURE 26. Integrated total transport ($m^3/sec \times 10^6$) based on continuity, July 1959.

CONCLUSIONS

1. The Alaskan Stream is the extension of the Alaska Current which rather than recirculating in the Gulf of Alaska continues westward along the Aleutian Islands and provides a source of dilute surface water, and relatively warm water (4°C) at depths of 300 m, to the northwestern North Pacific Ocean.

2. The Alaskan Stream is continuous as far westward as long. 170°E where it divides sending one branch into the Bering Sea and one southwestward which joins the eastward flowing Subarctic Current at about long. 165°E (Fig. 27). The Komandorski Ridge may influence the surface flow in this area.

3. Observed surface velocities along the shelf near Atka and Adak Islands are as high as 100 cm/sec. Relative speeds at the edge of the shelf are in excess of 30 cm/sec and mean values across the Alaskan Stream are about 10 cm/sec, referred to 1,000-m level.

4. Relative transport referred to the 1,000-m level is approximately 6×10^6 m^3/sec. A southward component of Ekman transport in the eastern passes during the period considered contributed to the confinement of the Alaskan Stream southward of the island chain as far westward as long. 175°E.

5. Less than 25 percent of the observed relative transport in the Alaskan Stream can be derived from the local wind-driven transport. Wind-driven transport integrated across lat. 50°N between long. 155°W and the North American coast amounts to 5.2×10^6 m^3/sec, essentially balancing the observed transport in the Alaskan Stream.

6. The concurrence of the observed transport with the transport calculated from wind stress supports the theory that the Alaskan Stream is driven primarily by the action of the wind over a broad region of the north-

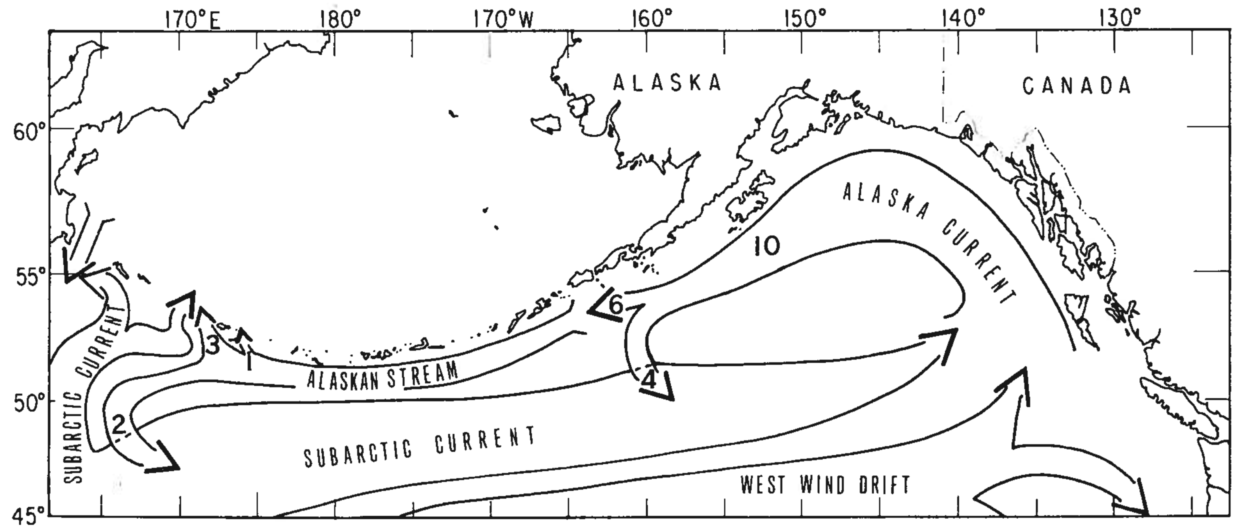

FIGURE 27. Schematic diagram of the Alaskan Stream. Numbers indicate volume transport ($m^3/sec \times 10^6$) referred to the 1000-m level.

east Pacific Ocean. The observed narrowness of the stream and continuity of transport also support the view that the Alaskan Stream is a western boundary current related to the general distribution of wind stress.

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