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Oceanography of the Subarctic Pacific Ocean

By Michitaka Uda[2]

Fisheries Research Board of Canada
Pacific Oceanographic Group, Nanaimo, B.C.

ABSTRACT

All available oceanographic data observed from 1951 through 1958, north of Lat. 35° N in the Pacific Ocean, the Sea of Okhotsk, and the Bering Sea, together with the relevant literature, were examined. There were sufficient data from 1955 through 1958 to prepare composite charts of salinity, temperature, dissolved oxygen and transparency of the near-surface waters in the Subarctic Region from Asia to America.

The Subarctic Pacific is defined as the region in which there is a marked halocline separating a low salinity upper zone from a more saline lower zone. It includes all areas having dichothermal temperature structure. Definite boundaries for the region are established. The Subarctic characteristics are limited to the upper zone and halocline. The lower zone characteristics are continuous from the Subtropic into the Subarctic.

From the data the principal features of structure of the properties have been shown, as well as the differences from one locality to another. The seasonal variations and also some non-seasonal variations have been discussed. The persistent current systems have been traced and their components named.

The processes for the maintenance of the structure by precipitation, evaporation, dissipation of upper zone water, and entrainment from the deep zone are discussed. A relation between the barometric pressure distribution and the transport in the non-persistent current systems is proposed.

CONTENTS

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[2]Author's present address: Tokyo University of Fisheries, 6 Chome, Shiba Kaigan Pori, Minato-ku, Tokyo, Japan.

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A. INTRODUCTION

THIS PAPER constitutes an attempt to present a concise but comprehensive review of the oceanography of what will be termed the Subarctic Pacific Oceanic Region. The term "region" will be considered to represent an oceanic area which is characterized generally throughout by one or more oceanographic features peculiar to the area. The Subarctic Pacific possesses features not found in any other region.

It is hoped that this report will serve a two-fold purpose: firstly, as a primer of the oceanographic knowledge that has been gained in the region up to the present time (1959) and secondly, as a guide for those whose interests are directed toward the various facets of the major fisheries of the area. With respect to the latter, a knowledge of the basic oceanographic features is a necessary prerequisite for any significant work on such aspects as the migration patterns and the feeding characteristics of commercially-significant fishes.

B. THE SUBARCTIC REGION

1. Geography

Figure 1 shows the Subarctic Region as defined in this paper, including all geographical terms used. The region involved is basically triangular in shape. At its northern apex (Lat. 66° N) is the narrow (31 miles or 58 km) and shallow (30 fathoms or 58 m) Bering Strait. To the south, the boundary is the northern limit of the Polar Front Region (page 127). This boundary lies on the Great Circle route from Tokyo to Los Angeles. In mid-ocean it lies generally between Lat. 40° and 45° N. The coast of North America forms the eastern boundary of the region. Along the western boundary lie the northern Japanese islands, the Okhotsk Sea and the Siberian coast. The Okhotsk Sea is roughly rectangular in shape, and is bounded on three sides by the Siberian coast (including the Kamchatka Peninsula). Access by water to the Okhotsk is possible only at the south side and is limited by the Kurile Island Chain. Access to the Bering Sea is influenced in a similar manner by the Aleutian Island Chain, the southern boundary of that sea.

2. Bathymetry

It is to be noted that this and the following section have drawn heavily upon information given by Fleming (1955). This information is included in the present work for the sake of completeness.

The bathymetry of the region (Fig. 1) is notable for the marked contrasts in depth. The greatest depths, those exceeding 4000 fathoms (about 7300 m), are found in deep trenches that lie parallel to the island chains. By contrast, the northern and eastern parts of the Bering Sea, and much of the northern Okhotsk Sea, overlie the continental shelf. The shelf is represented approximately by the area between the coast and the 100-fathom (about 200 m) contour. These parts of the Bering Sea are extremely shallow, possessing a maximum depth of only about 30 fathoms (55 m). Basins having depths of about 1750 fathoms (3200 m), intermediate to the extremes just noted, are present in both the Okhotsk and southern Bering Seas. The remainder of the region is essentially an abyssal plain, averaging about 2500 fathoms (about 4500 m). In the northeastern Subarctic, several chains of seamounts are present. Some of the seamounts rise to within 400 fathoms (about 730 m) of the surface.

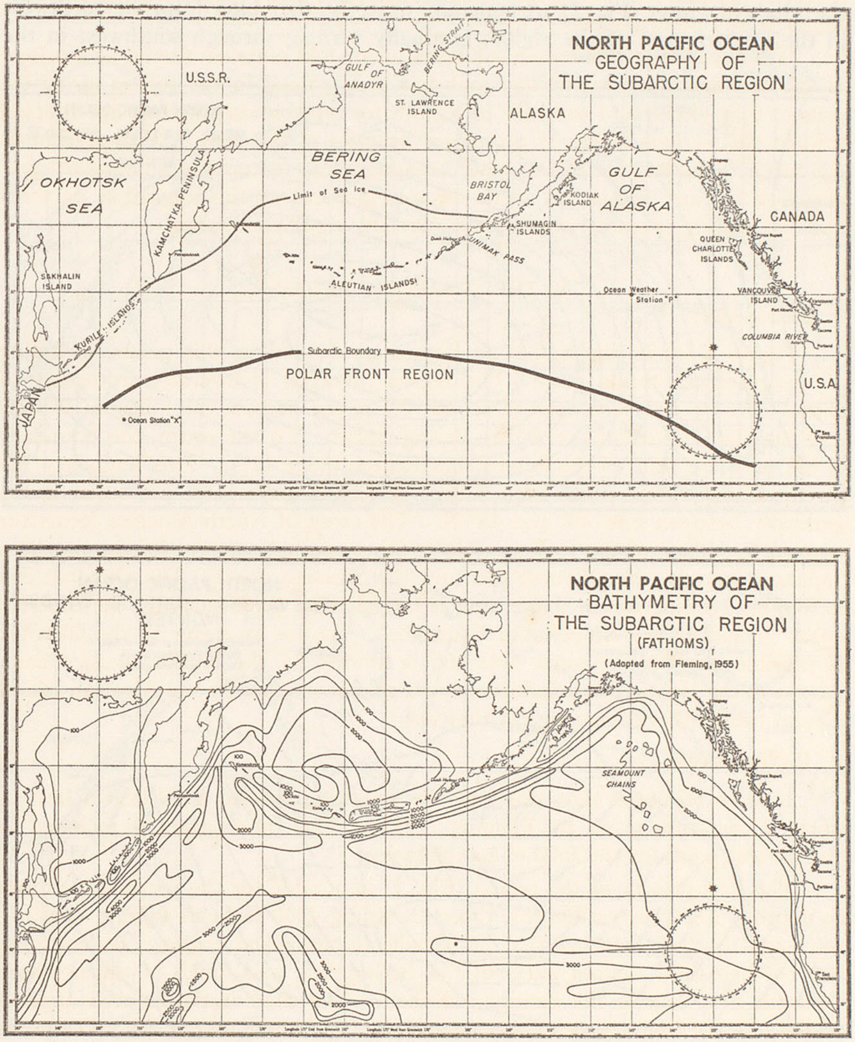


FIG. 1. Geography and Bathymetry (fathoms) of the Subarctic Region (adapted from Fleming, 1955).

3. Climatology

(i). Winds. During the winter (December to early April) the Subarctic Pacific Region is usually under the influence of a low-barometric pressure cell centred in the area of the Bering Sea and the Aleutian Islands (the Aleutian Low) (Fig. 2). Coincident with this, two high-pressure cells (the Siberian and the North American Arctic High) occur. Under the influence of these cells the winds in winter (Fig. 2) are generally westerly; they blow from the northwest in the western part of the region, gradually turning through southwest in the

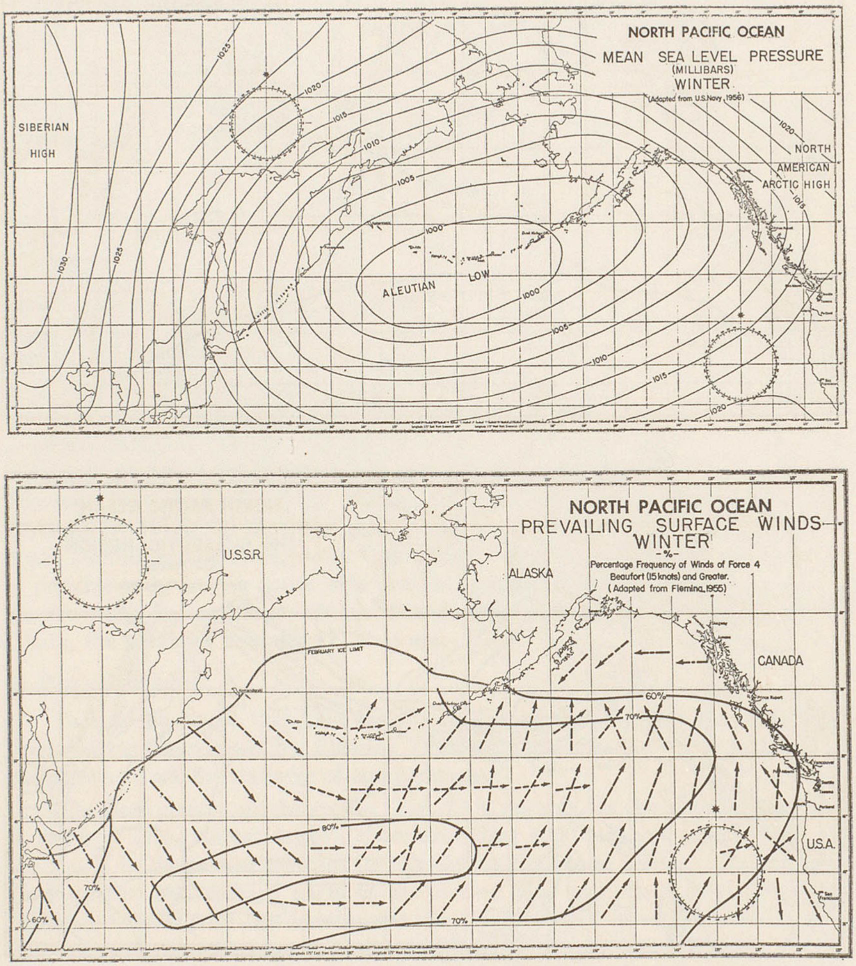


Fig. 2. Mean Sea Level Pressure (millibars), Winter (adapted from U.S. Navy, 1956) and Prevailing Surface Winds, Winter (adapted from Fleming, 1955).

central and eastern Pacific. In the northern part of the Gulf of Alaska, easterly winds prevail.

In early summer (Fig. 3) the Aleutian Low vanishes and the North Pacific High, centred between about Lat. 30° and 40° N, becomes predominant. The prevailing westerlies of the winter months are then replaced by the predominately south or southwest winds (Fig. 3). These warm southerly winds cause extensive

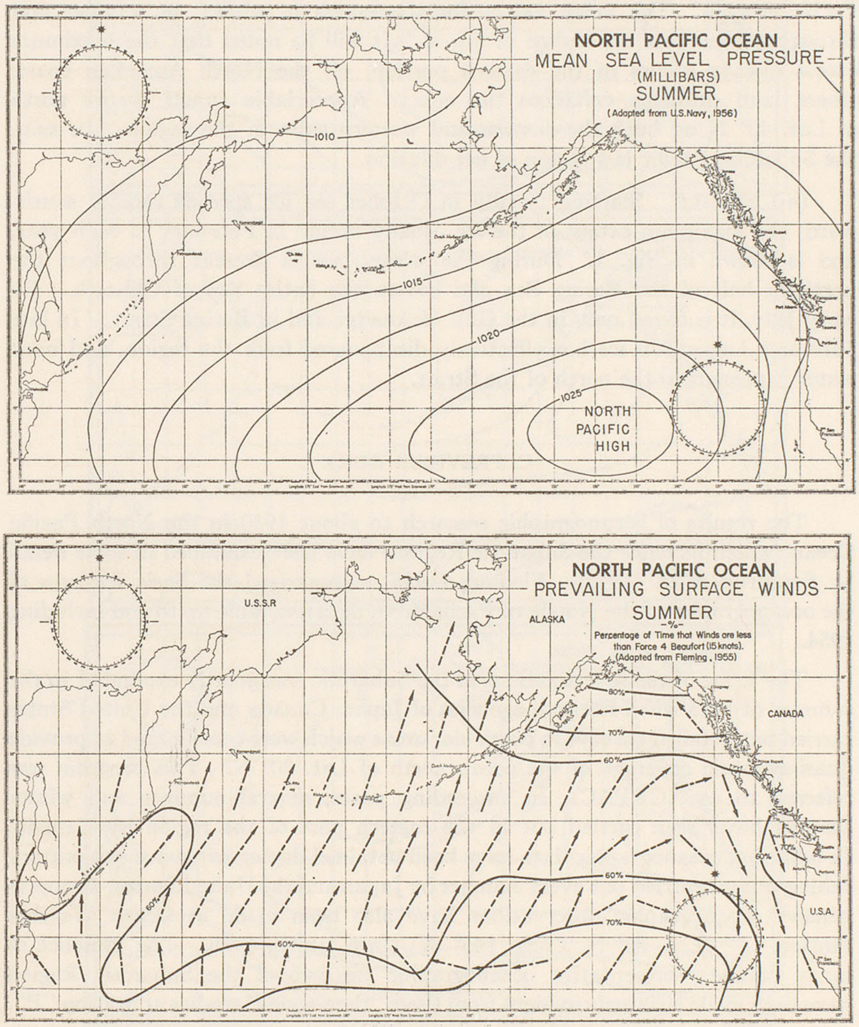


FIG. 3. Mean Sea Level Pressure (millibars), Summer (adapted from U.S. Navy, 1956) and Prevailing Surface Winds, Summer (adapted from Fleming, 1955).

"advection" fogs during their passage over the cold waters off Kamchatka and the Aleutian Chain. On the eastern side of the Subarctic Region (off the Canadian and U.S. coasts) the dominant summer winds are from a northwest direction. Light variable winds occur in the Gulf of Alaska at this time.

(ii). PRECIPITATION. It has been shown (Jacobs, 1951) that precipitation exceeds evaporation throughout the year over essentially the entire Subarctic Pacific Region. The mean values for the excess in winter, in summer and throughout the year are shown in Fig. 4. It will be noted that the maximum excess always occurs in the eastern portion, off the North American coast, where land drainage enhances the effect. Appreciable runoff occurs north of Lat. 45° N on both the eastern and western sides of the region. In sum, the Subarctic Pacific is a region of net dilution.

(iii) SEA ICE. Starting usually in October sea ice spreads rapidly southward; the maximum extent of the ice usually occurs in February of each year, and is shown in Fig. 1. During the winter, ice is present throughout the northern half of the Bering Sea and covers the entire Sea of Okhotsk. By early July, it is found only in the Gulf of Anadyr and in Bering Strait. In late July and August, all ice has effectively disappeared from the region, and open water is present to the north of the Strait.

C. PREVIOUS WORK

The results of oceanographic research to about 1940 in the North Pacific Ocean (which includes the Subarctic Region) have been examined in some detail by Sverdrup *et al.* (1942). Fleming (1955) summarized the basic features of the oceanography of the Northern Pacific from data available up to and including 1954.

The scope of the investigations in the Subarctic was greatly expanded in the summer of 1955 when research agencies of Japan, Canada and the United States carried out detailed surveys in particular areas which were coordinated to provide quasi-synoptic coverage of the ocean north of Lat. 20° N. This program was referred to as NORPAC. In succeeding years, several summer and winter surveys have been carried out in the eastern part of the region by Canada. In addition, oceanographic data have been obtained during extensive exploratory fishing cruises carried out every summer by Japan and the United States. Comprehensive oceanographic observations have also been made at Ocean Weather Station "P" (Lat. 50° N, Long. 145° W) since late in 1956. The presence of many of the representative oceanographic features of the Subarctic Region have been made strikingly evident from these "time-series" studies at Station "P".

It is on the data obtained during and subsequent to Project NORPAC that this paper is primarily based. The sources of the data are appended.

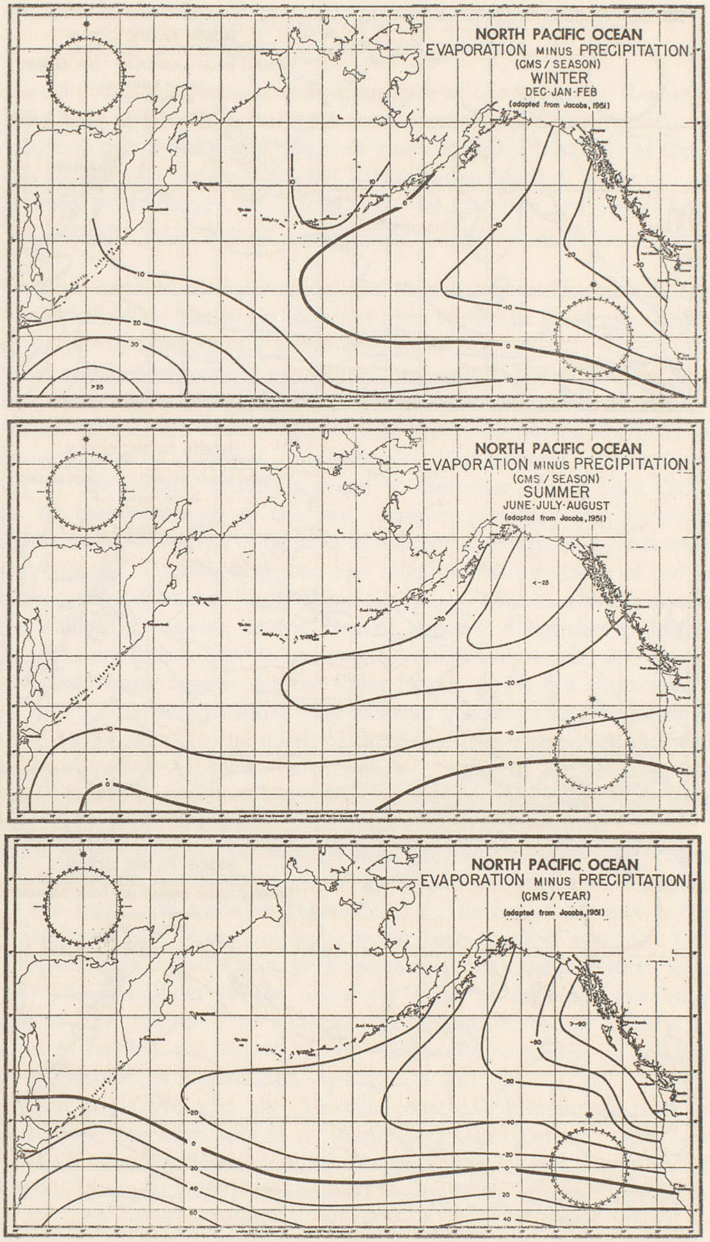


FIG. 4. Evaporation minus precipitation, winter (cms/season), summer (cms/season), and annual (cms/year) (adapted from Jacobs, 1951).

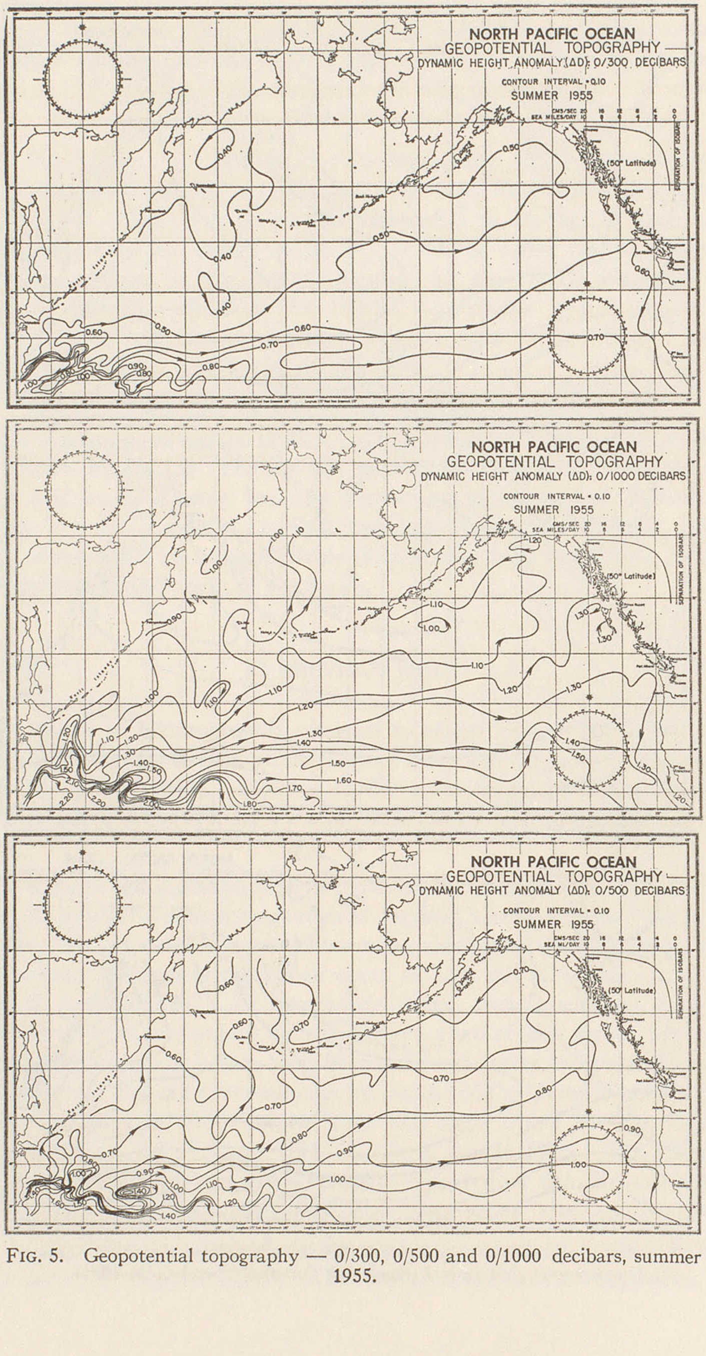


FIG. 5. Geopotential topography — 0/300, 0/500 and 0/1000 decibars, summer 1955.

D. THE PERSISTENT CURRENT SYSTEM IN THE SUBARCTIC PACIFIC REGION

Figure 5 shows the geopotential topography of the Subarctic Region referred to the 300-, 500-, and 1000-decibar surfaces respectively. The data were collected during the summer of 1955 (the period of most intensive coverage of the region). Also, the topography referred to the 1000-decibar surface is shown for each summer from 1956 through 1958 (Fig. 6). The currents are presumed to flow along the isobaths with a speed inversely proportional to the spacing between isobaths.

The more important components of the persistent circulation in the upper layers in the Subarctic Pacific (as deduced from recurrent features in these charts of geopotential topography) are shown and named in Fig. 7.

A cold southward-moving current of low salinity, the Oyashio, is present along the eastern side of the Kurile Islands and along the northern Japanese island (Uda, 1938a, b). The warm, saline Kuroshio moves northeastward off the Japanese islands to about Lat. 40° N. This current is relatively swift, especially at the surface, having a speed of the order of one knot (50 cm/sec) or more. At the confluence of the two currents, extensive mixing occurs both horizontally and vertically. The area is usually characterized by a widespread system of eddies. The mixed water, and most of the remainder of the two currents, turn eastward as the "West Wind Drift". This has a much slower speed than the Kuroshio, moving at the rate of about 2–4 nautical miles/day (4–8 cm/sec). The speed is relatively uniform to a depth of at least 300 m.

The mixed water constitutes the Polar Front, which is a transitional region about 2° to 4° of latitude in width; the location varies across the Pacific (Kawai, 1955; Uda, 1938a, b). To the north of the Front, the water is termed Subarctic (the region of primary interest to this paper), while to the south it is termed Subtropic. In the Front there is a marked gradation of oceanographic properties, from those typical of the one region to those typical of the other.

Off the American coast, the West Wind Drift (which has become progressively more dilute because of precipitation) divides. In the region of separation, the currents are weak and variable (Doe, 1955). The Subtropic water, the Polar Front water, and part of the Subarctic water turn south to form the California Current which flows toward the tropics and eventually joins the North Equatorial Current. The remainder of the Subarctic water is deflected northward and circulates around the Gulf of Alaska. The dynamic centre of this cyclonic circulation is termed the Alaska Gyre. This centre varies in position from 200–300 miles south of Kodiak Island (Fig. 1, 5 and 6).

Between the Gyre and the Alaskan coast the circulation both narrows and accelerates, reaching speeds of about 4–6 miles per day (8–12 cm/sec). From Kodiak Island to Unimak Pass the current is termed the Alaska Stream (Schott, 1935; Bennett, 1959), and beyond Unimak, the Alaska Stream Extension (page 168). The water is relatively warm and, especially in summer, becomes progressively more diluted as it moves westward. The near-surface layers finally enter the Bering Sea through the Aleutian passes nearest the Alaskan

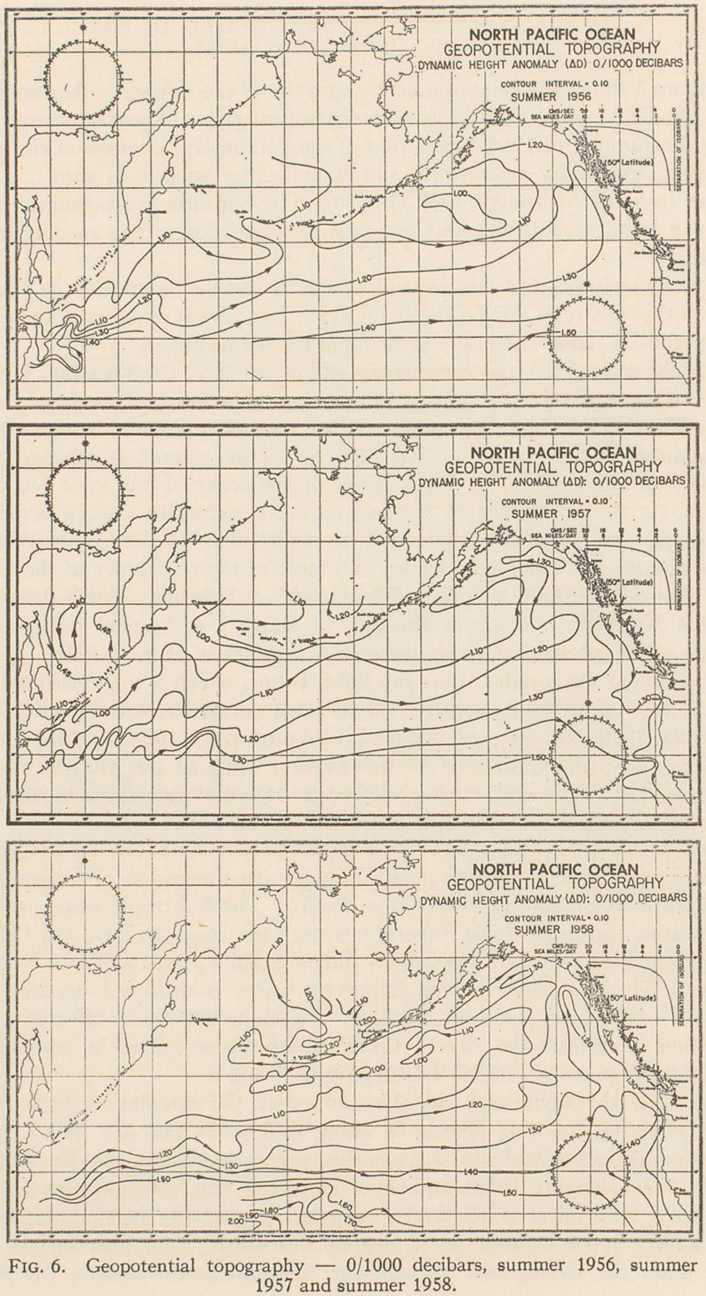


FIG. 6. Geopotential topography — 0/1000 decibars, summer 1956, summer 1957 and summer 1958.

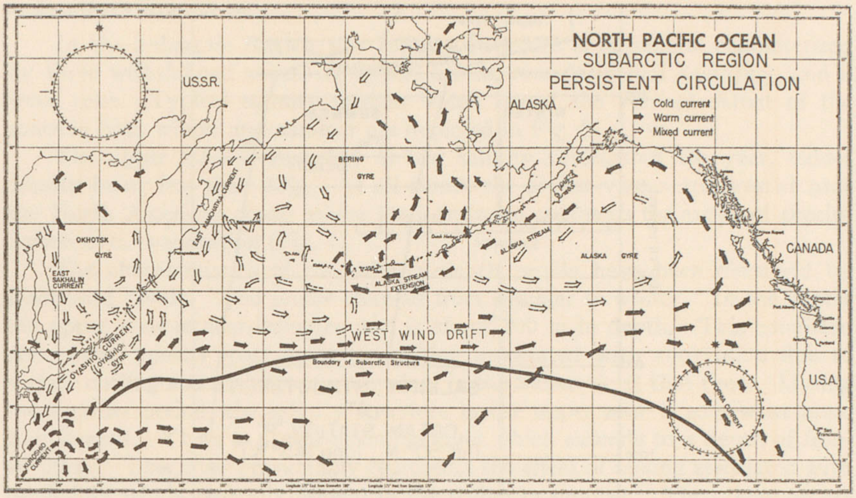


FIG. 7. Persistent circulation in the Subarctic Region.

mainland. The deeper waters move through the deeper passages which lie farther to the west.

The water in the Bering Sea moves in a cyclonic path. Some passes east of St. Lawrence Island and moves through Bering Strait. Observations in the Strait indicate that the flow through it is predominantly northward (Goodman *et al.*, 1942). Of the remainder, which turns southward along the Siberian coast, one part (modified by climatic conditions in the immediate area) becomes the East Kamchatka Current. The other part is absorbed into the cyclonic Bering Gyre which appears to be present in the Bering Sea.

In turn, part of the East Kamchatka Current moves into the Okhotsk Sea through passages in the North Kurile Island Chain, is modified, and circulates in a cyclonic path. On the western side of the sea it becomes the East Sakhalin Current. Most of this current moves out through the middle Kurile passages to form, with the East Kamchatka Current, part of the cold Oyashio (page 167). A portion remains in the (cyclonic) Okhotsk Gyre.

The circulation in the Subarctic Pacific is closed, aside from the water lost from the system through the Bering Strait to the Arctic Ocean, and through the California Current to the Subtropic Region. It appears that about 6 years are necessary for the round trip (Tully and Barber, 1960).

This picture of the persistent currents has been deduced from summer data. Confirmation of its major features has come, in the eastern part of the region at least, from geopotential topography derived from winter data (Dodimead, 1958) and from the results of drift-bottle releases (Dodimead and Hollister, 1958).

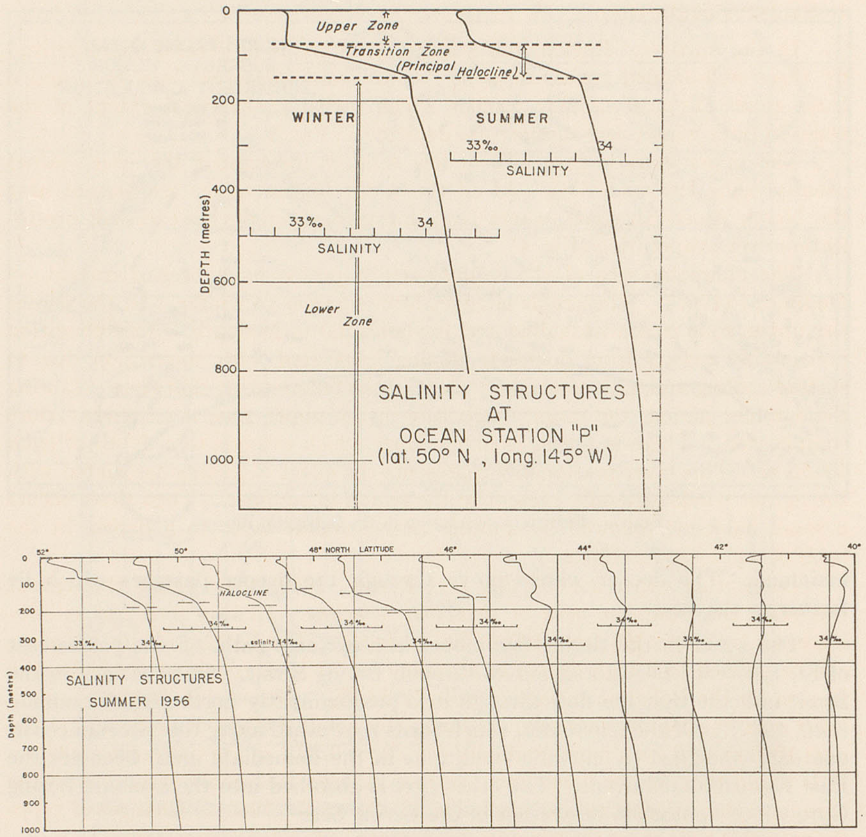


FIG. 8. Salinity structures at Ocean Station "P", winter and summer and in a north–south section in mid-Pacific Ocean.

E. SALINITY OF SUBARCTIC WATERS

It has been found that the most effective method of defining Subarctic waters is by a description of the *structure*, which term here denotes the variation of seawater properties in a *vertical* column or section. On the basis of structure (primarily that of salinity, but also to a lesser degree, that of temperature) the north Pacific can be divided into three distinctly separate regions. Each region is characterized by a different structure, each structure in turn possessing the same general characteristics throughout a region. The three are — in order of decreasing latitude — the Subarctic, the Polar Front, and the Subtropic (Fig. 1). As befits the title of this paper, the structure found in the first will be discussed in detail. A few salient features of the remaining regions will be noted, for purposes of general comparison.

1. Salinity Structure

In the Subarctic Region, the salinity structure is in general characterized by three well-defined zones, an *upper*, a *transition* (*principal halocline*), and a *lower* zone. Typical salinity profiles which reveal the vertical extent of the zones in both winter and summer are depicted in Fig. 8.

The upper zone is generally of low salinity throughout the year, values usually being less than 33‰. At all times, the lowest values are present near the North American coast, where land drainage augments the excess of precipitation over evaporation (Fig. 4).

The characteristics of the salinity–depth profile have been discussed by Dodimead (1961). The profile changes from summer to winter. In the winter the upper zone is near-isohaline and is about 100 m in depth. This condition presumably results from complete mixing associated with convection due to surface cooling and with storms prevalent in the Subarctic at that time. During the summer period, the salinity structure in the upper zone is somewhat more complicated. There is a near-isohaline layer which extends to a depth of from 10–30 m. This layer presumably represents the effect of mixing associated with the lighter winds generally present in the region in summer. Then there occurs a small halocline followed by another near-isohaline layer to a depth, in the open ocean, of about 100 m.

Below this upper zone (Tully and Barber, 1960) there exists throughout the year, the *transition zone* or *principal halocline*. In this zone the salinity increases by about 1‰ from 100 to 200 m depth. The bottom of this halocline can be defined to a high degree of accuracy (±0.1‰) throughout the Subarctic Region by the isohaline surface of salinity 33.8‰. Below this, in the so-called *lower zone*, the salinity increases gradually (by about 0.6‰) but uniformly in the next 1000 m, and even more gradually (by about 0.2‰) in the succeeding 1000 m. It is found that, in the lower zone, surface-induced (seasonal) variations in the oceanographic properties are negligible (Tully *et al.*, 1960). Changes of properties in this zone must therefore be due primarily to advective effects.

Dodimead (1961) has shown that the "three-zone" structure is characteristic of the mid-ocean portion of the Subarctic Region. However, it is not present in all coastal areas. In localities where intense mixing occurs because of topographical features, such as in the passes of the Aleutian Island Chain, the upper boundary of the halocline may be as deep as 400 m and the lower boundary may be indefinite. This represents the greatest depth at which the halocline has been found. It is shallowest in summer in areas such as the Bering and Okhotsk Seas, where sea ice is formed in winter. No halocline is present in winter in the shallow (i.e. the northern and eastern) portions of the Bering Sea. Convective overturn associated with sea-ice formation renders the water isohaline from surface to bottom. Isohaline conditions are present effectively throughout the year near the southern edge of the continental shelf in the Bering Sea (Fig. 1). These conditions may have their origin in the mixing occurring over the shallow bottom in that area. The lower diagram of Fig. 8 (Dodimead, 1961) illustrates the sequence of structures along a north–south section between 165° and 175°

west longitude. This section extends from the Subarctic through the Polar Front into the Subtropic Region in mid ocean.

The term "Polar Front" usually has the connotation of a zonal belt in which there is a marked meridional temperature gradient. This occurs generally between 40° and 45° N in the north Pacific Ocean. There is a corresponding marked change of salinity through this belt, at least in mid-ocean; hence the term Polar Front is used here to designate the transition belt in both salinity and temperature structures.

Southward through the Polar Front, the halocline becomes progressively less distinct (Dodimead, 1961). At the southern side of the belt, the structure becomes effectively isohaline to a depth of several hundred metres. Surface salinities increase to about 33.8‰ through the Front to the southern edge. As many as three salinity inversions may be found in the upper 150 m or so, but temperature considerations (page 147) indicate that these can be density-stable. Considerable fluctuations in salinity structure, both in space and time, occur in this transition belt.

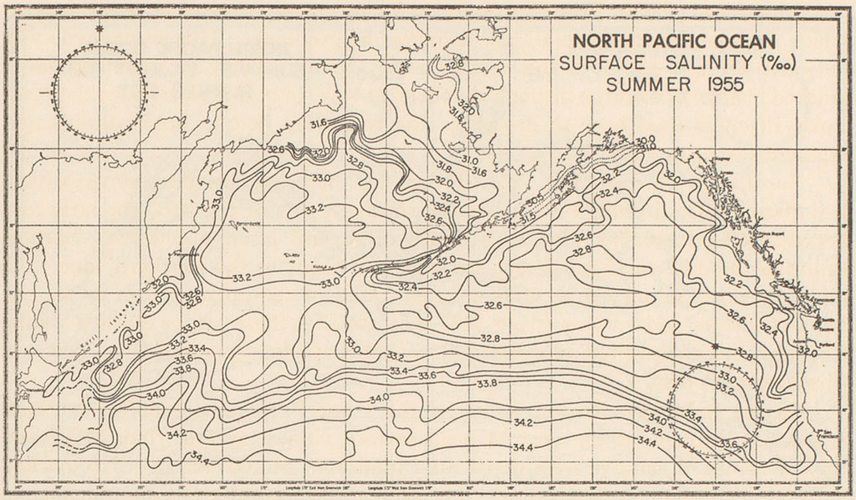
In the Subtropic Region, south of the Polar Front, the salinity structure is marked throughout the year by high surface values (greater than 34‰) and by a decrease of salinity with depth to a minimum at 300–800 m. No marked halocline is present in this region. The high surface salinity is a consequence of the excessive evaporation (Fig. 4) due to the prevailing Westerlies (Fig. 2 and 3).

The salinity and structure in the deep water, below the salinity minimum and the halocline, are similar and continuous from the Subtropic through the Subarctic.

2. Horizontal Distribution of Salinity

Summer and winter horizontal distributions of surface-layer salinities, during the years 1955 through 1958 inclusive are shown, to the extent of the data available, in Fig. 9 through 13. It appears that distributions at the surface and at 10 m depth in the Subarctic both present the same basic features (compare Fig. 9, 10 with 11, 12). It is believed, however, that conditions at 10 m are most nearly representative of the prevailing seasonal state. Presumably those at the surface would be more influenced by transient effects, (evaporation, dilution, etc.) especially in the summer periods. In winter, conditions at the two levels would presumably be practically identical, because of the strong convection and wind mixing. It was found possible to draw similar conclusions with respect to both temperature and dissolved oxygen distributions. Therefore, throughout the remainder of this paper, surface-layer conditions will be represented by the data from 10 m depth.

In the surface-layer salinity pattern, several permanent features are noticeable. The trend of the isohalines appears to be basically east–west throughout the year. Marked modifications, however, occur along the shoreline and are attributable to the current systems, coastal winds and land drainage. A permanent salinity maximum is present in the Gulf of Alaska, south of Kodiak Island. Maxima have also been observed to occur in both the southern Bering



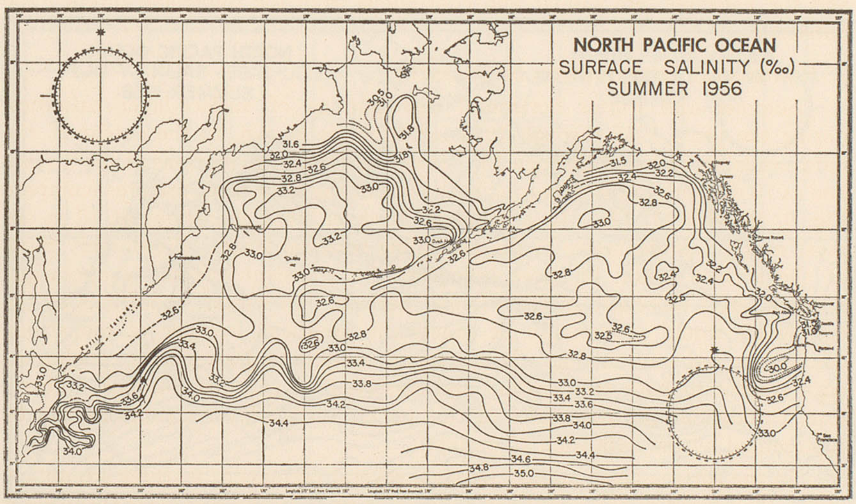


FIG. 9. Surface salinity, summer 1955 and summer 1956.

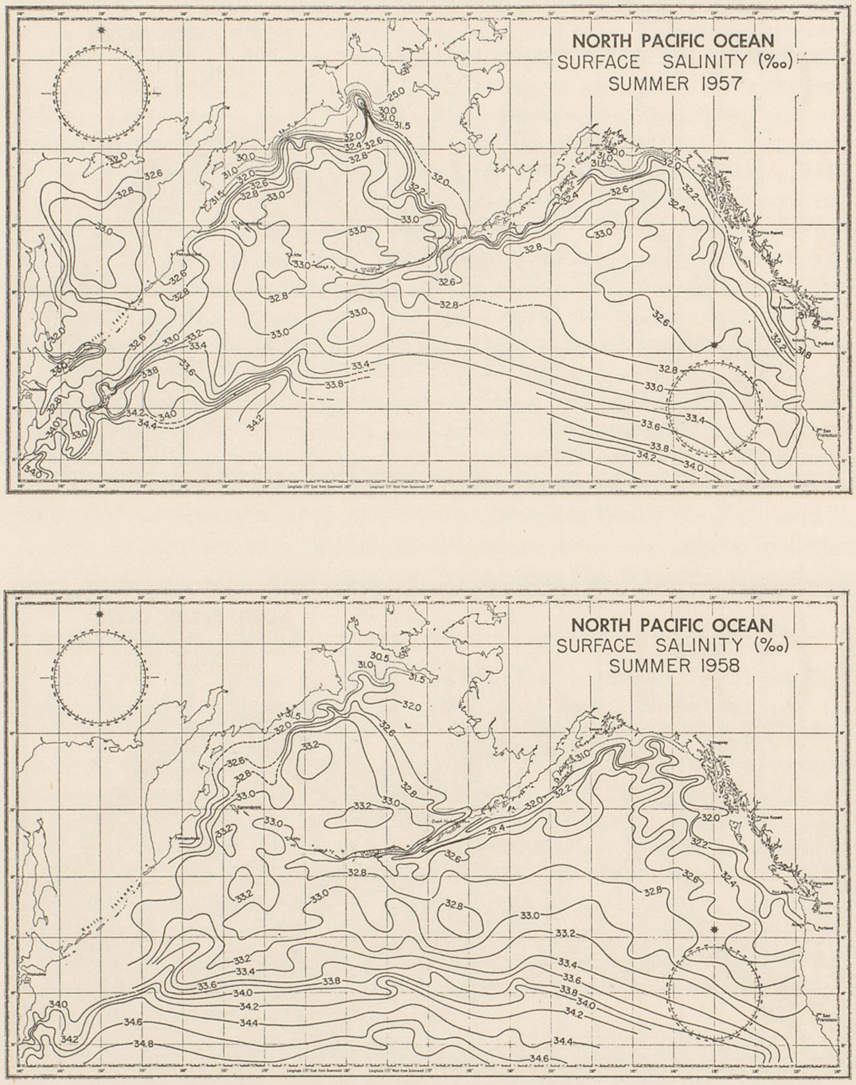
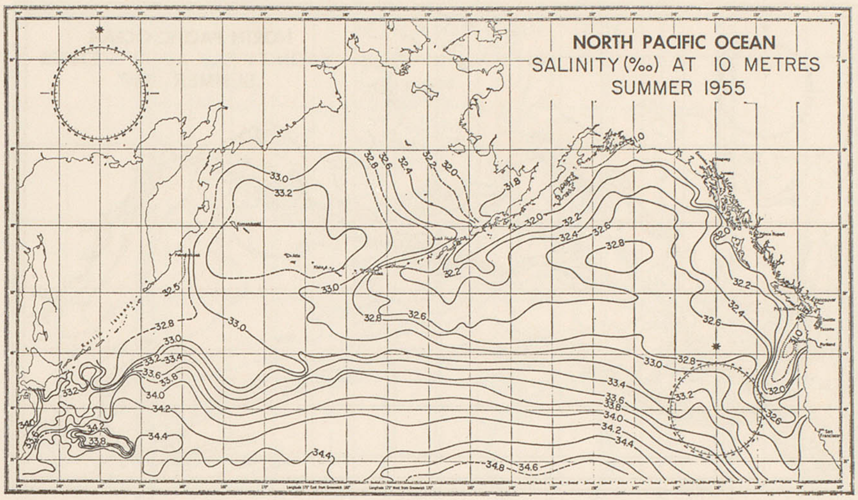


FIG. 10. Surface salinity, summer 1957 and summer 1958.



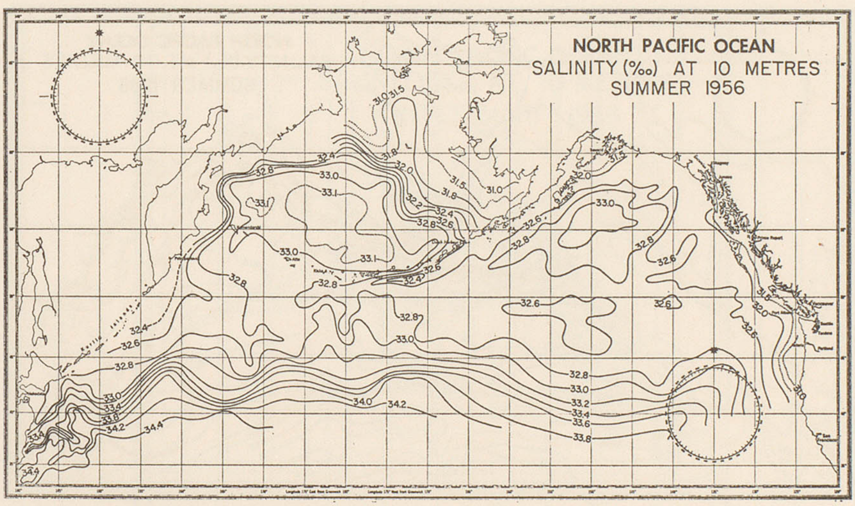


FIG. 11. Salinity at 10 metres, summer 1955 and summer 1956.

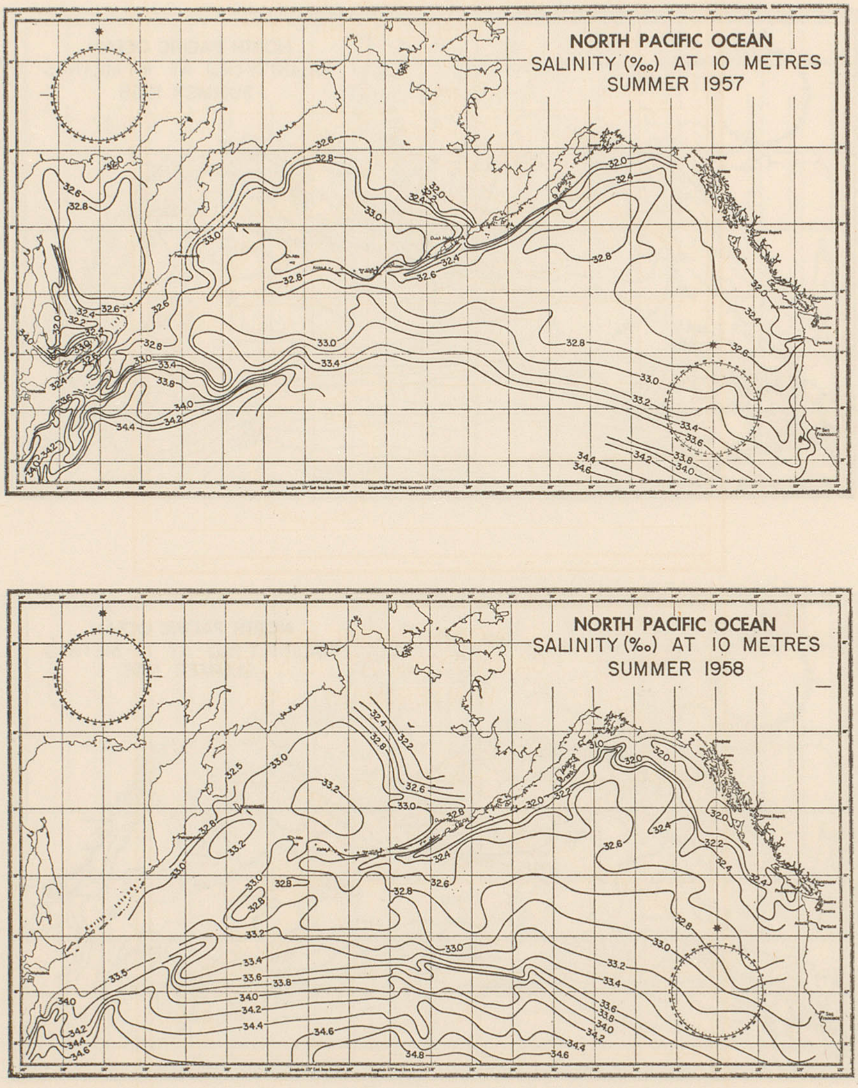


FIG. 12. Salinity at 10 metres, summer 1957 and summer 1958.

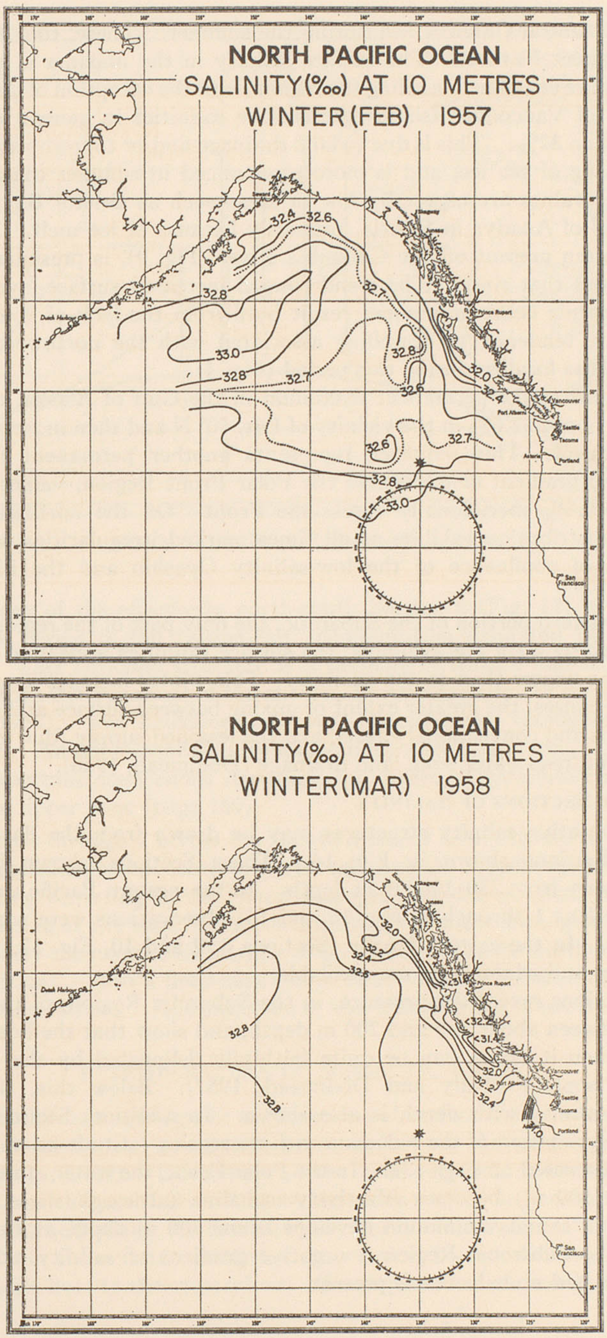


FIG. 13. Salinity at 10 metres, winter (Feb) 1957 and winter (March) 1958.

Sea and the central Okhotsk Sea during the summer. These, too, are believed to be permanent features. The surface salinity in the maxima is about 33‰. Along the entire coastline bounding the region (with the exception of the American coast south of Vancouver Island) the surface salinities in general decrease to values less than 32‰. This is due to land drainage and/or to fresh water released by the melting of sea ice, and is more pronounced in summer than in winter. The summer values are especially low in areas such as Bristol Bay in Alaska, and the Gulf of Anadyr in Siberia, primarily because of ice melt. A localized minimum, often present off the Columbia River (Fig. 9), is presumably due to discharge from that river. The trend toward increased surface-layer salinities off the California coast in summer result both from the lack of land drainage, and from the tendency to upwelling associated with the northwesterly winds prevalent in the locality during this period (Fig. 3).

Southward of the permanent maximum in the Gulf of Alaska, the salinity decreases to about 32.6‰ in the vicinity of Lat. 50° N and then increases through the Polar Front. This sequence represents another permanent feature. A fairly marked gradient characterizes the Polar Front Region, values increasing as much as 0.8‰ meridionally across the Front. Off the northern Japanese islands the distribution exhibits at all times marked irregularities which occur in the area of confluence of the low-salinity Oyashio and the high-salinity Kuroshio.

In the eastern portion of the Subarctic, the only part of the region for which significant winter data are available, the surface salinities appear to be slightly greater in the winter than in the summer. This winter increase is attributable to two main causes; the greater extent of mixing between surface and deep water due to storms and convectional effects, and the lessened supply (due to snow and ice storage) of fresh water from land drainage (Dodimead, 1958).

3. Vertical Sections of Salinity

Representative salinity structures may be drawn from the data observed along the Sections shown in Fig. 14. These Sections, shown in Fig. 15 through 19 vary from 250–1000 m in depth. In the western Pacific and Aleutian regions (Sections 1 through 7, Fig. 15 and 16) observations were made only in the summer. In the eastern Pacific (Sections 8, 9 and 10, Fig. 17, 18 and 19) both summer and winter data are available.

The sections reveal the presence, in the Subarctic Region, of the principal halocline between about 100 and 200 m depth, and show that the bottom of the halocline, when it exists, can be quite faithfully delineated by the position of the 33.8‰ isopleth (Tully and Dodimead, 1957). Below this, the gradual increase of salinity with depth is also shown. In addition, Sections 1 and 8 reveal the degradation of the halocline with decreasing latitude as was shown in Fig. 8 and discussed on page 132. In the Polar Front, the water at intermediate depths (200–500 m) becomes effectively isohaline (although slight inversions may occur); a salinity minimum develops below 300 m depth at the southern limit. In the Subtropic Region a negative gradient of salinity exists in the upper layers, and no halocline is present.

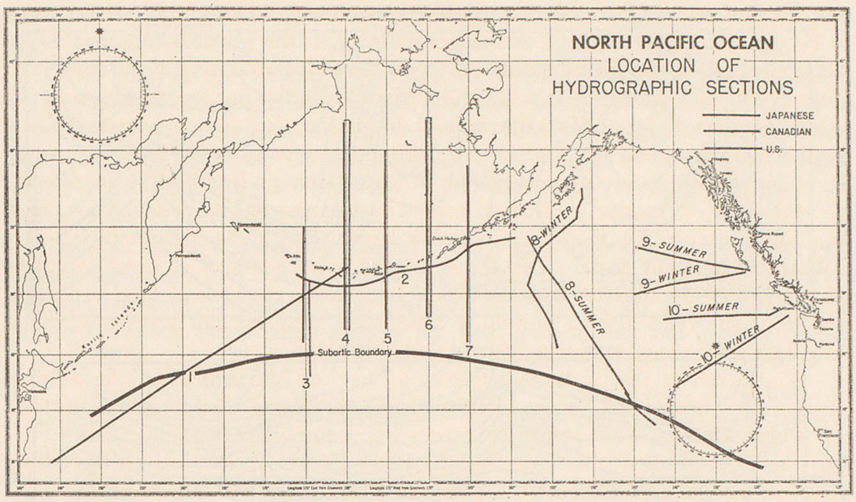


FIG. 14. Location of Hydrographic Sections.

A feature of the effectively north-south Section 8 (Fig. 17) in the eastern part of the region is the "dome-like" configuration exhibited by the deeper isohalines. This anticline structure is a permanent feature and is presumably associated with the dynamics of the anti-clockwise circulation of the Alaska Gyre (Bennett, 1959). The halocline, however, is continuous throughout the area of the dome. An upper-zone salinity maximum occurs in this area. Its existence leads to important conclusions regarding the means of replenishment of saline water in the upper zone (page 169).

Sections 3 through 7 across the Aleutian Island Chain reveal the three-zone structure typically present south of the islands, and the deepening of the halocline by the mixing occurring in the island passages. Strong gradient currents and intensive mixing at the edge of the continental shelf in the Bering Sea (page 120) are revealed by the marked inclination of the isohalines at the northern extremity of these sections. Also a deep-zone dome structure was present in 1956 south of the Aleutian Islands. It is unknown at this time whether it too is a permanent feature.

F. TEMPERATURE OF SUBARCTIC WATERS

Amounts of heat, giving rise to variations of surface temperature significant on even a diurnal scale, are gained from, or lost to, the surroundings by the oceans (Tabata, 1960). Effectively all the heat supplied to the oceans comes originally from the sun. Over 85% of the sun's radiation penetrating the sea surface is absorbed within the first 10 m of clear oceanic water. In turn, heat is dissipated from the oceans, mainly by evaporation, back radiation, and conduction of sensible heat to the atmosphere. These processes are modified in turn by

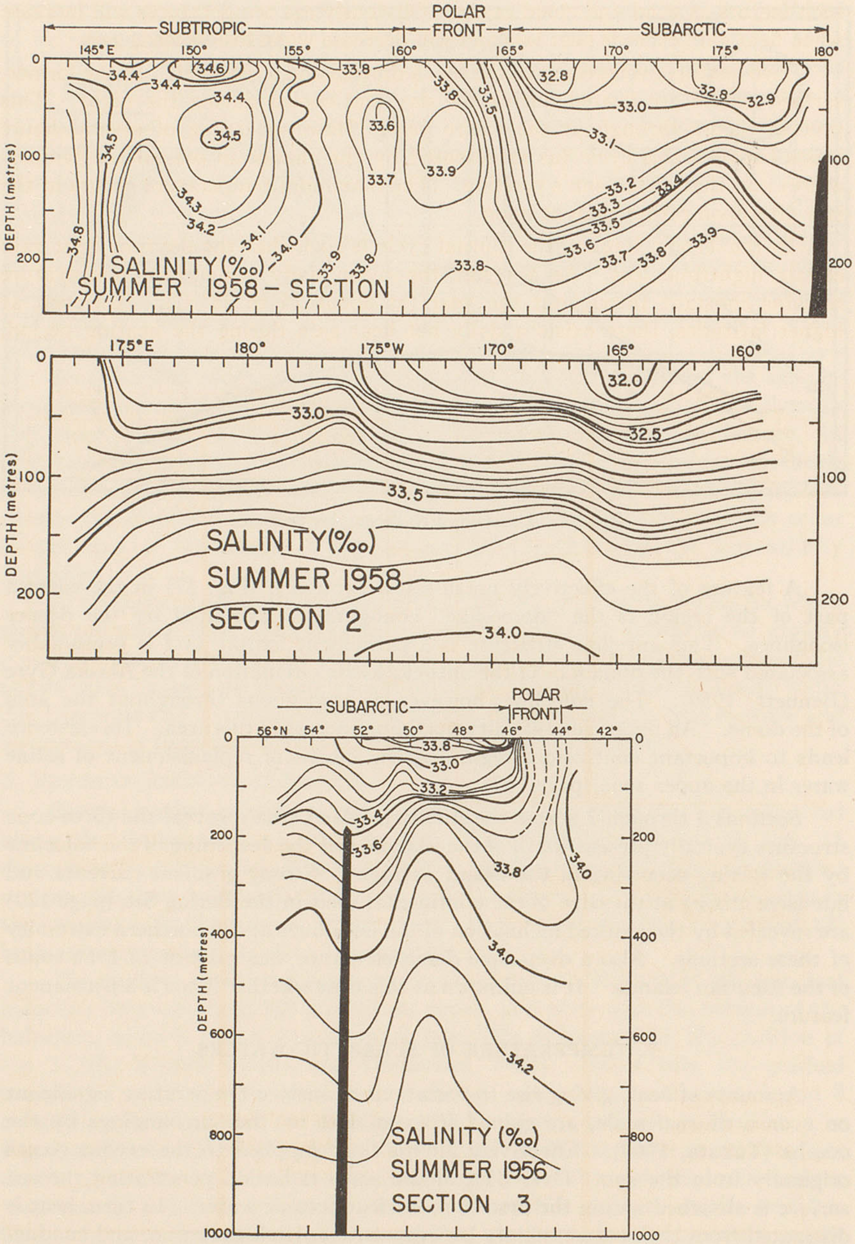


FIG. 15. Salinity, Section 1, summer 1958, Section 2, summer 1958 and Section 3, summer 1956.

such factors as wind and cloud cover. Obviously sea temperatures will increase when heat gain exceeds heat loss and will decrease when loss exceeds gain.

Two heating cycles occur in the sea: a diurnal and a seasonal. In the former, surface waters are heated during the day and cooled during the night. This process occurs throughout the entire year. However, a lag of several hours exists between this cycle (as represented by the change in sea-surface temperature) and that of the sun's position. The maximum temperature occurs in the late afternoon rather than at noon.

In the Tropic Region, the diurnal cycle is such that the daytime heat gain equals nighttime loss. As a result the mean daily sea-surface temperature remains constant throughout the year (at about 25 to 28° C). However, at higher latitudes, there exists a daily net heat gain during the heating period,

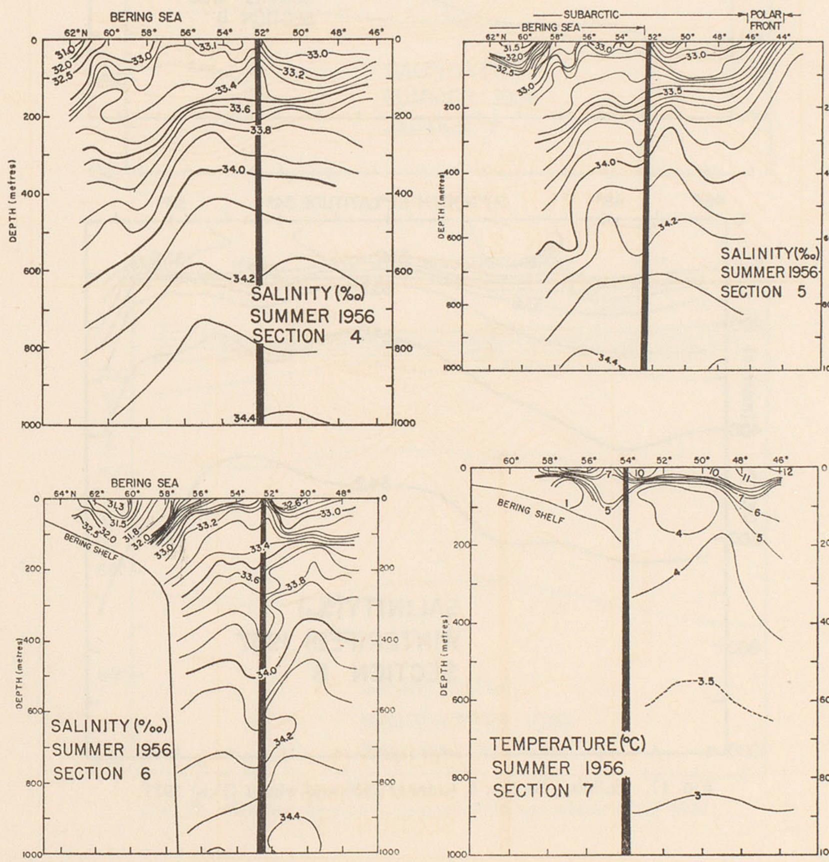


FIG. 16. Salinity, summer 1956, Sections 4, 5, 6 and 7.

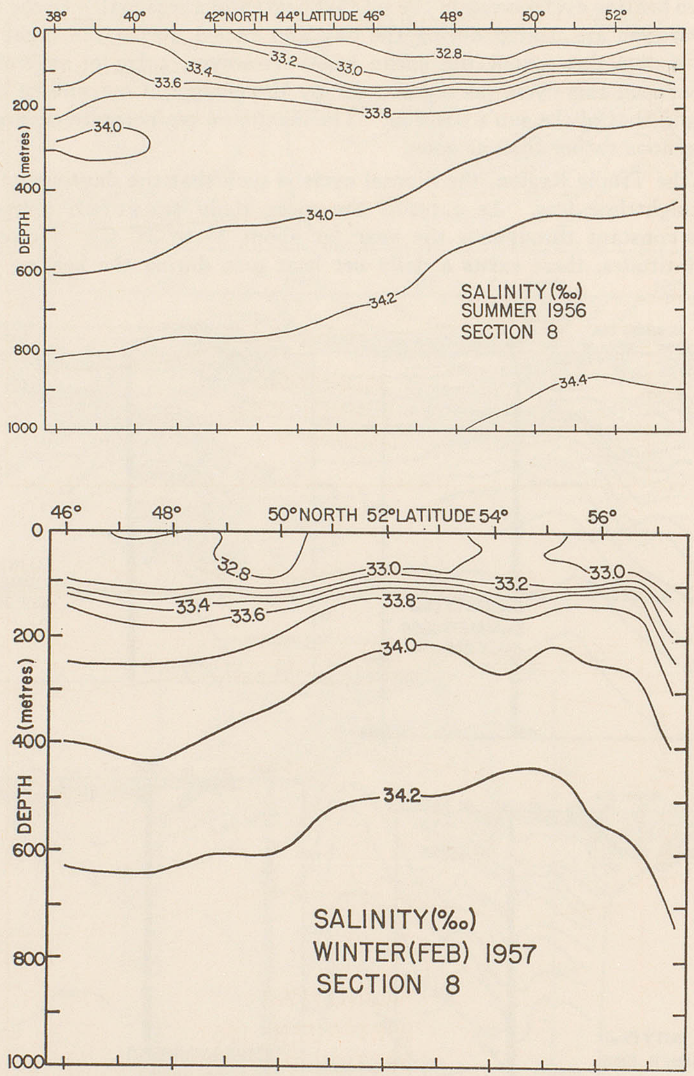


FIG. 17. Salinity, Section 8, summer 1956 and winter (Feb) 1957.

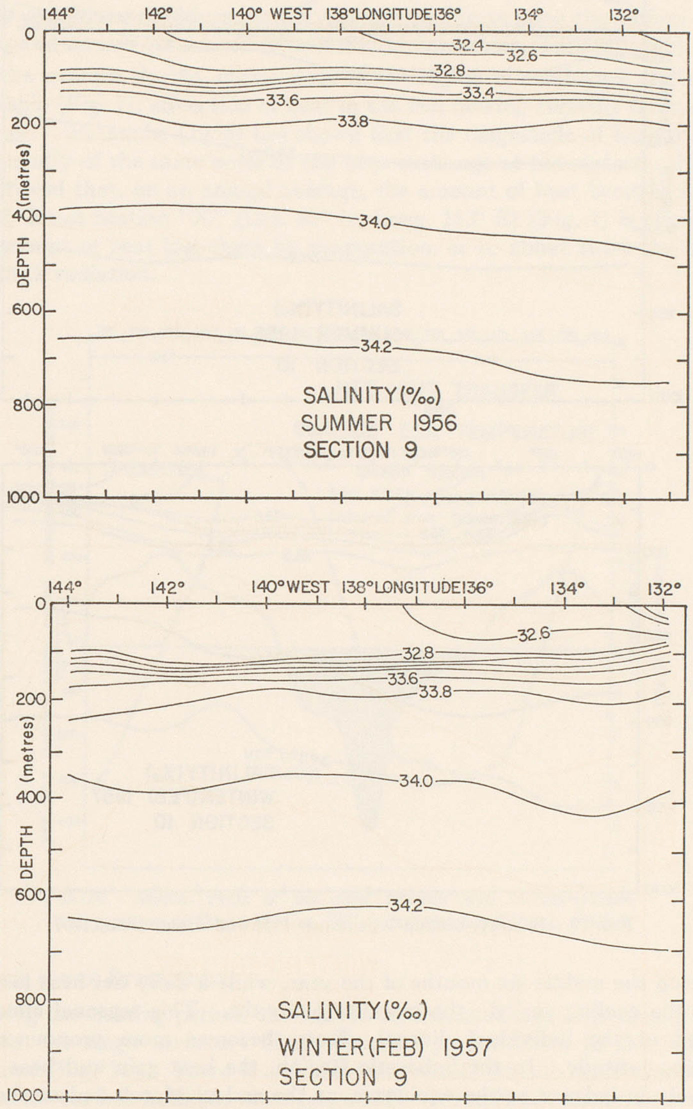


FIG. 18. Salinity, Section 9, summer 1956 and winter (Feb) 1957.

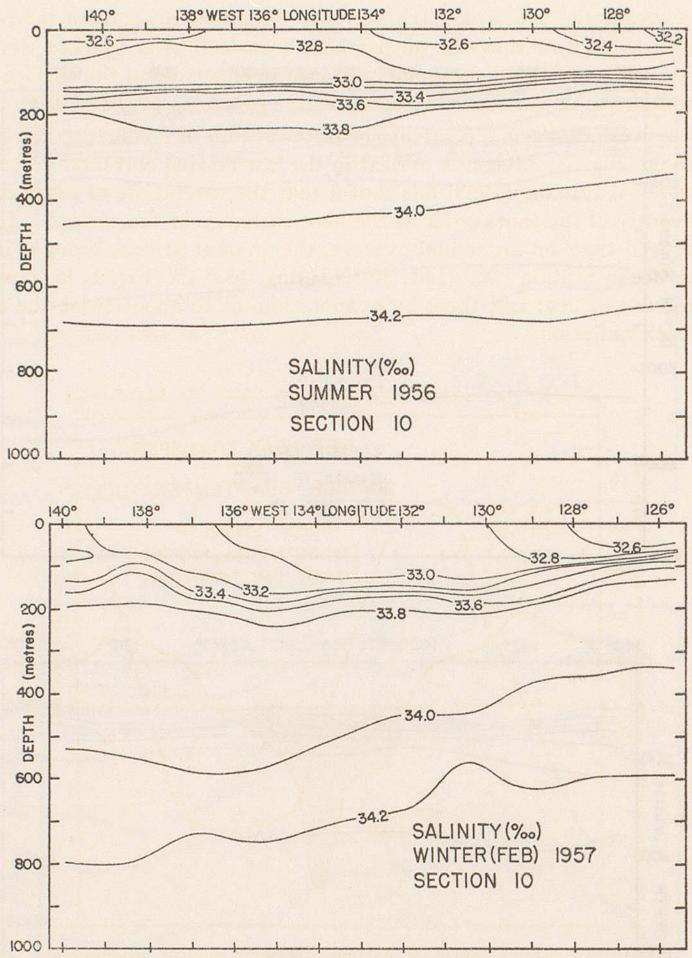


FIG. 19. Salinity, Section 10, summer 1956 and winter (Feb) 1957.

i.e. during the middle six months of the year, while a daily net heat loss occurs during the cooling period—the remaining months. This seasonal effect, compounded of the individual diurnal effects, becomes more pronounced with increasing latitude. In the Subarctic Region, the heat gain and heat loss are effectively in balance at the equinoxes, at the end of March and at the end of September. Fig. 20 shows, for the Subarctic, both a typical annual (seasonal) cycle of the net heat transfer across the sea–atmosphere interface and the accompanying sea-surface temperature cycle. The data were obtained at Ocean Weather Station "P" (Lat. 50° N, Long. 145° W) (Fig. 1). It is seen that, in the open ocean, the times of maximum and minimum sea-surface temperature coincide with the two times of no net heat transfer. Any marked deviations from this conditions would presumably be due to effects upon the surface waters by horizontal and/or vertical eddy or advective processes (Tabata, 1960).

In the western Pacific, particularly in the region of confluence of Kuroshio and Oyashio (Fig. 7), advection of heat in the fast flowing currents is a considerable factor. Watanabe (1955) has shown that the magnitude of heat transport was frequently of the same order as the heat exchange at the surface. Koizumi (1956) proved that, on an annual average, the amount of heat brought into the region of Ocean Station "X" (Lat. 39° N, Long. 153° E) (Fig. 1) is comparable to the amount of heat lost there by evaporation, or to about twice the amount lost by back radiation.

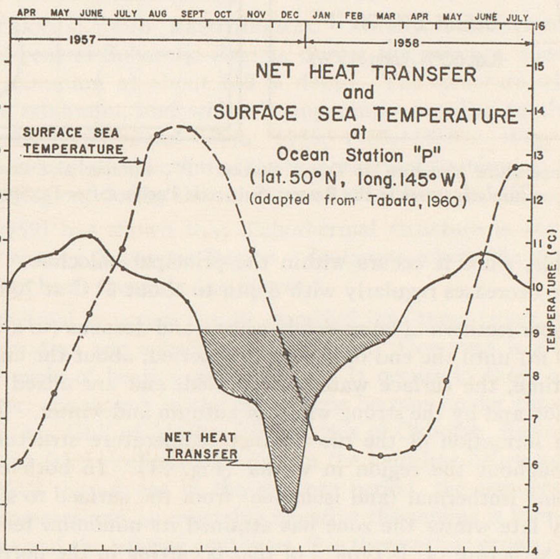


FIG. 20. Annual cycle of net heat transfer and of sea-surface temperature at Ocean Station "P" (adapted from Tabata, 1960).

1. Temperature Structure

During the heating period, the heat is accumulated and is gradually mixed downward. A single structure of temperature predominates in the Subarctic at this time (Fig. 21). There is a shallow isothermal surface layer which corresponds to the surface isohaline layer (page 131). It results, similarly, from mixing by the light winds prevailing during the period. Below the surface layer, a sharp negative thermocline is present. Below this, there follow in turn another shallow isothermal layer and, over much of the area, a positive thermocline. The latter extends to a depth of about 150 m. This temperature inversion is

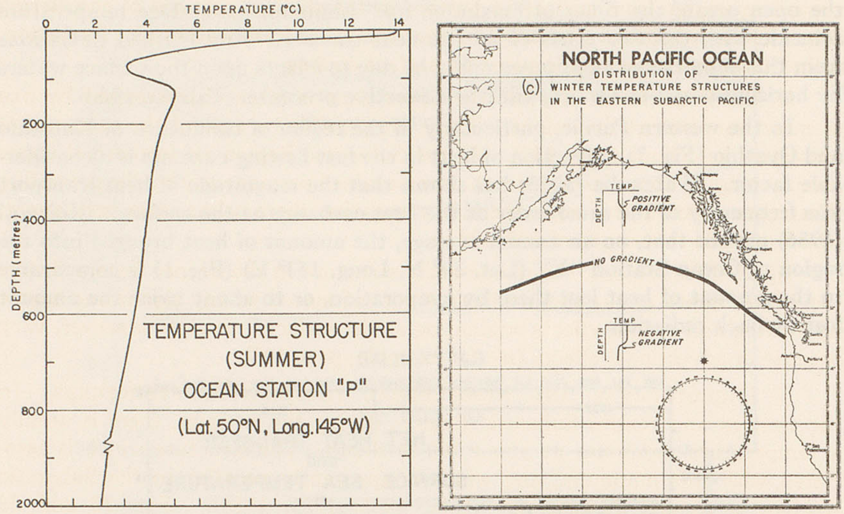


FIG. 21. Temperature structure at Ocean Station "P", summer and distribution of winter temperature structures in the Eastern Subarctic Pacific (after Dodimead, 1958).

density-stable, since it occurs within the principal halocline. Below this, the temperature decreases regularly with depth to about 2° C at 2000 m.

The upper isothermal layer slowly warms and deepens (to a maximum depth of about 30 m) until the end of the heating period, about the end of September. After this time, the surface waters are cooled, and are mixed downward both by convection and by the strong winds of autumn and winter. These conditions lead to the formation of the *two* distinct temperature structures which occur widely throughout the region in winter (Fig. 21). In both cases, the upper layers become isothermal (and isohaline) from the surface to a depth of about 100 m. By late winter the zone has attained its minimum temperature of the year. The structure (a) is typical of that occurring in the northern part of the region. The profile indicates that a positive gradient (thermocline) exists below the isothermal layer. The warmer water at this depth is density-stable because the thermocline occurs within the principal halocline. In this case, the cooling (net heat loss) is great enough so that the upper 100 m of water become colder than the deeper water. By contrast, the structure in the southern part of the region is still characterized by a negative thermocline. This, however, is much less pronounced than the summer thermocline. Here the surface cooling is insufficient to render the surface layers colder than those at depth. The distribution of these two structures throughout the eastern Subarctic is shown in the latter diagram of Fig. 21. Northward and southward of the boundary between the two types of structure, the positive and negative thermoclines become more pronounced. Isothermal conditions, arising as a result of convective overturn, prevail in winter from surface to bottom in the shallow portions of the Bering Sea. These conditions are caused by the removal of fresh water in ice formation, rather than by convection induced by surface cooling.

In the Polar Front Region, the upper zone is often characterized by one or more temperature inversions. These are possible because of the existence of associated haloclines (page 132). In the Subtropic Region an isothermal upper zone, averaging about 20 m in depth, is maintained permanently by the prevailing Westerlies. Below this zone there is a negative thermocline, also permanent, which extends to a depth of 500–600 m.

In all three regions, the temperature structure in the deeper waters (below the winter depth of the thermocline) is not influenced to any significant degree by heating and cooling effects originating at the surface. It may, however, vary because of advectional effects (see for example, page 148).

(i) Dichothermal and mesothermal temperature structures. In the western and central Subarctic Region, during the summer, there is usually a temperature minimum at about 100 m depth. The structure which is associated with this minimum, and which is apparently peculiar to the Subarctic Region, is termed *dichothermal* (Uda, 1935, 1955, 1958). Fig. 22 and 23 illustrate the horizontal distribution of the dichothermal (minimum) temperature in the Subarctic Region, during the summers of 1955 through 1958.

Bennett (1959) has shown that dichothermal structure is not continuous throughout the eastern Subarctic Region. Instances where the structure does *not* occur have *not* been shown in the figures.

The dichothermal structure can be classified into three types, according to the mode of formation and subsequent behaviour. One type is a consequence of the seasonal cycle of heating and cooling. It occurs in those areas where, during the winter, the waters in the upper zone become colder than the waters in the halocline (e.g. Gulf of Alaska, Fig. 21, Sea of Okhotsk, Bering Sea, etc.). A *positive* gradient (thermocline) is formed in the halocline below an isothermal upper zone. With the advent of the summer period, the temperature of the surface layers increases and an overlying negative thermoline is formed. As the period progresses and the heating continues, this thermocline grows and deepens. In this type of dichothermal structure there is an isothermal "temperature-minimum" layer situated between the upper negative and the lower positive thermocline. The layer generally occurs just above the halocline (60–100 m depth). The characteristic temperature ($T_{minimum}$) changes very little during the course of the summer period (Dodimead, 1961). During the following period of net cooling the upper thermocline is progressively erased; the dichothermal structure vanishes and is replaced by the winter structure typical of the area.

A second type of dichothermal structure originates, by the same process, in the areas of sea-ice production, chiefly the Okhotsk Sea. The cold, more saline iso-layer remaining after the formation of the ice extends to depths which are generally within the shallow and narrow halocline present in this sea. The temperature of this layer is between –1.8 and –1° C. The layer (in the Okhotsk

Sea) follows the persistent circulation (page 129) and moves through the middle Kurile passages into the western Pacific. Because of mixing along its path, especially in the island passages, the layer becomes warmer (to about 1° C). A portion of the layer underruns the upper layers of the warm (to 20° C) saline (to 35‰) Kuroshio to form the greatest part of the Oyashio Undercurrent. The resulting temperature minimum at depth represents the second type of

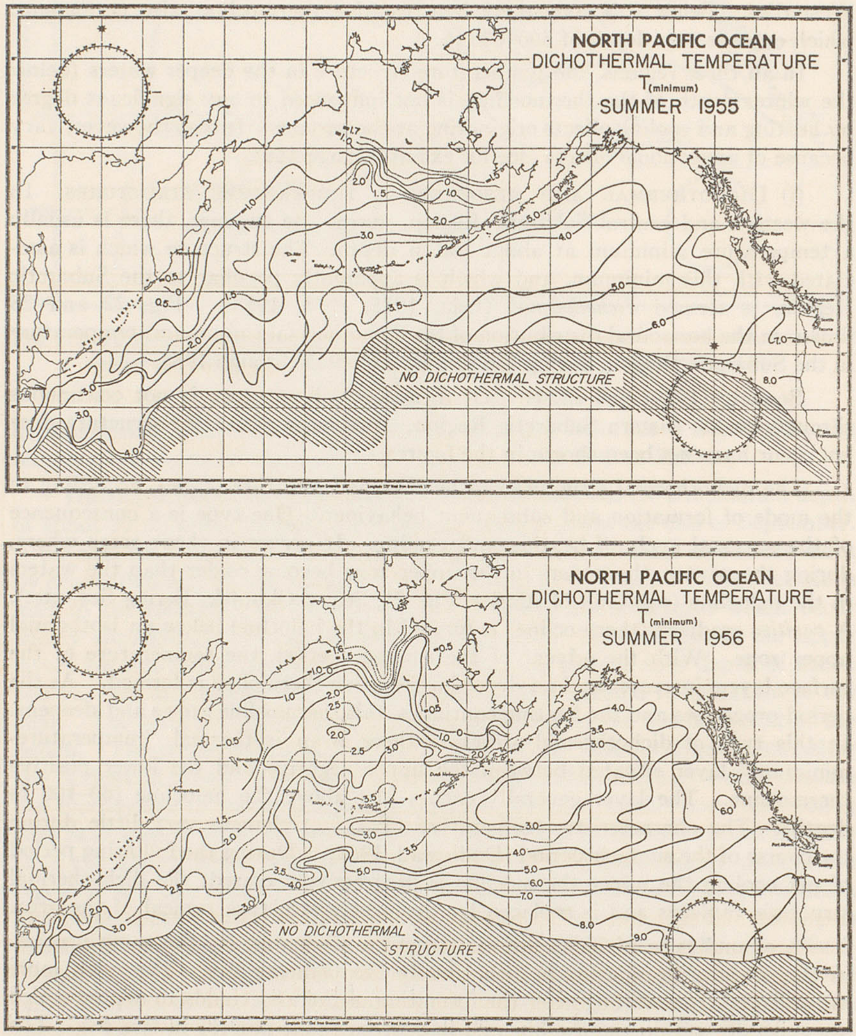


FIG. 22. Dichothermal temperature, summer 1955 and summer 1956.

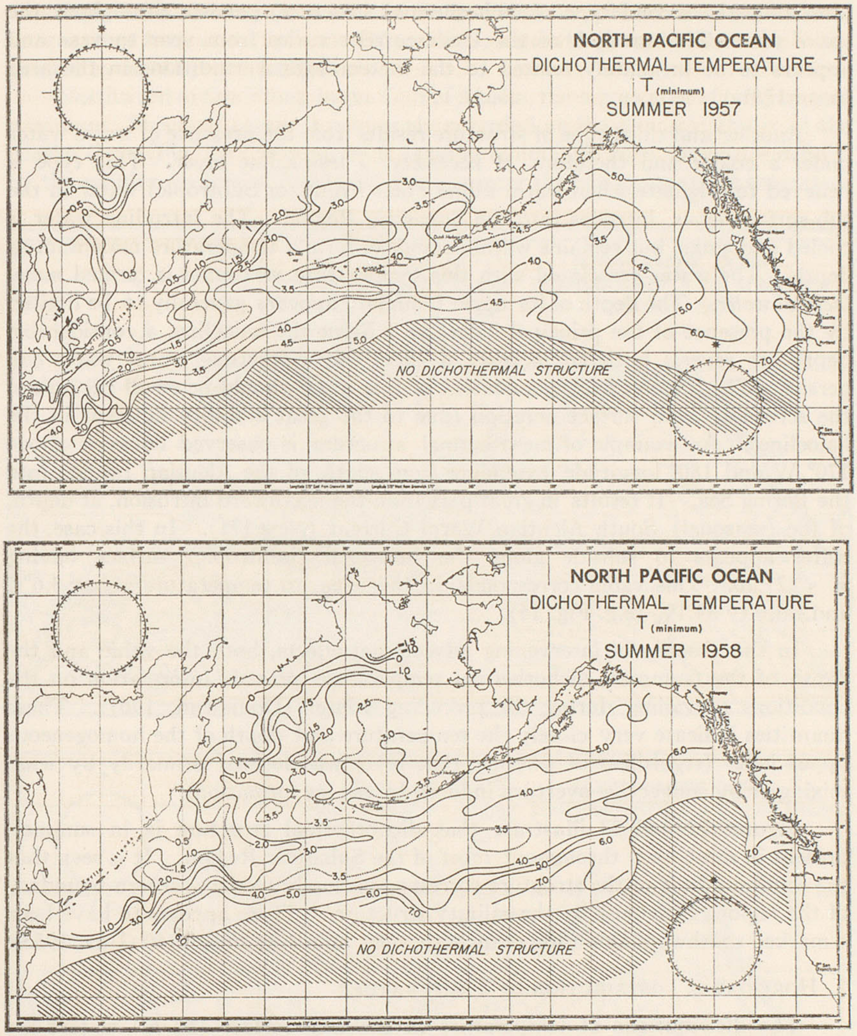


FIG. 23. Dichothermal temperature, summer 1957 and summer 1958.

dichothermal structure. The passage at depth of the water through the Kurile passages and its (modified) continuation as the Undercurrent can be "traced" by the horizontal distribution of dichothermal temperature as shown, for example, in Fig. 22. It may be noted that the progress of the Undercurrent (which is relatively fresh compared to Kuroshio water) can also be followed by means of a salinity minimum at depth. The outflow through the middle Kurile passes (and thus the second type of dichothermal structure) persists throughout the *entire year*. The strength of the Undercurrent varies from year to year and appears to be intimately related to the meteorological conditions in the area (page 173).

Another and third type of structure results from the presence of warm water under a colder and therefore, of necessity, a less saline layer. This type is believed to originate when warm saline Polar Front (or Subtropic) water in the sub-surface layer, intrudes into the Subarctic Region. The intruding water is cooled and sinks, but remains warm enough to form a temperature maximum at depth. The water associated with this temperature *maximum* is termed *mesothermal water*. The depth of its upper boundary appears generally to be limited by the presence of the principal halocline. *Below* the halocline, a considerable depth (to several hundred metres) of effectively isothermal warm water can persist. The dissolved oxygen content of this water should be small because of the screening from surface aeration (due to the great stability inherent in the halocline). An example of mesothermal structure is observed between about 170° W and 180° longitude extending from south of the Aleutian Islands into the Bering Sea. It results in great part from the northward intrusion, at depth, of the (seasonal) South Aleutian Warm Current (page 171). In this case, the water appears to subside along the isopycnal (isentropic) surface having σ_t = 27.20. This value corresponds, in this area, to temperature about 3.6°C and salinity 34.1‰ (e.g. Fig. 16).

In the absence of intervening advectional effects, both the value and the depth of the (summer) dichothermal temperature provide information on the conditions prevailing during the preceding winter (Dodimead, 1961). These quantities indicate very closely the temperature and depth of the homogeneous upper layer (regardless of whether that layer was formed primarily by wind mixing or by convective overturn induced by sea-ice formation).

Figures 22 and 23 illustrate that dichothermal structure is, in summer, present in some form throughout most of the Subarctic Region. It is seen that the southern limit of the structure corresponds closely to the southern boundary of the region, as defined by the salinity structure. There appears to have been a marked southward extension of dichothermal structure since 1940 (Uda, 1955).

2. Horizontal Distribution of Temperature

The distributions of temperature in the near-surface (10 m) waters are shown in Fig. 24, 25 and 26. They correspond to the salinity data shown in Fig. 11, 12 and 13.

During the summer the general configuration of the isotherms is east–west, except near the continental margins. The winter data are sparse and limited to the Gulf of Alaska. However, they indicate that the pattern is similar to that of summer, although the water is about 7 C° colder. This is confirmed by the monthly mean sea-surface temperature charts published by the U.S. Weather Bureau (1938).

Figures 24, 25 and 26 show the Polar Front as a meridional gradient of temperature from about latitude 35° to 45° N, somewhat broader than the Front defined by salinity (page 132).

Northward of this Front, in the Gulf of Alaska, there is an area of temperature minimum. This cold centre is commonly regarded as the dynamic centre of the Alaska Gyre (Fig. 7). In summer it appears to coincide with the centre defined by the salinity maximum (page 139) but in winter the two appear to be separated.

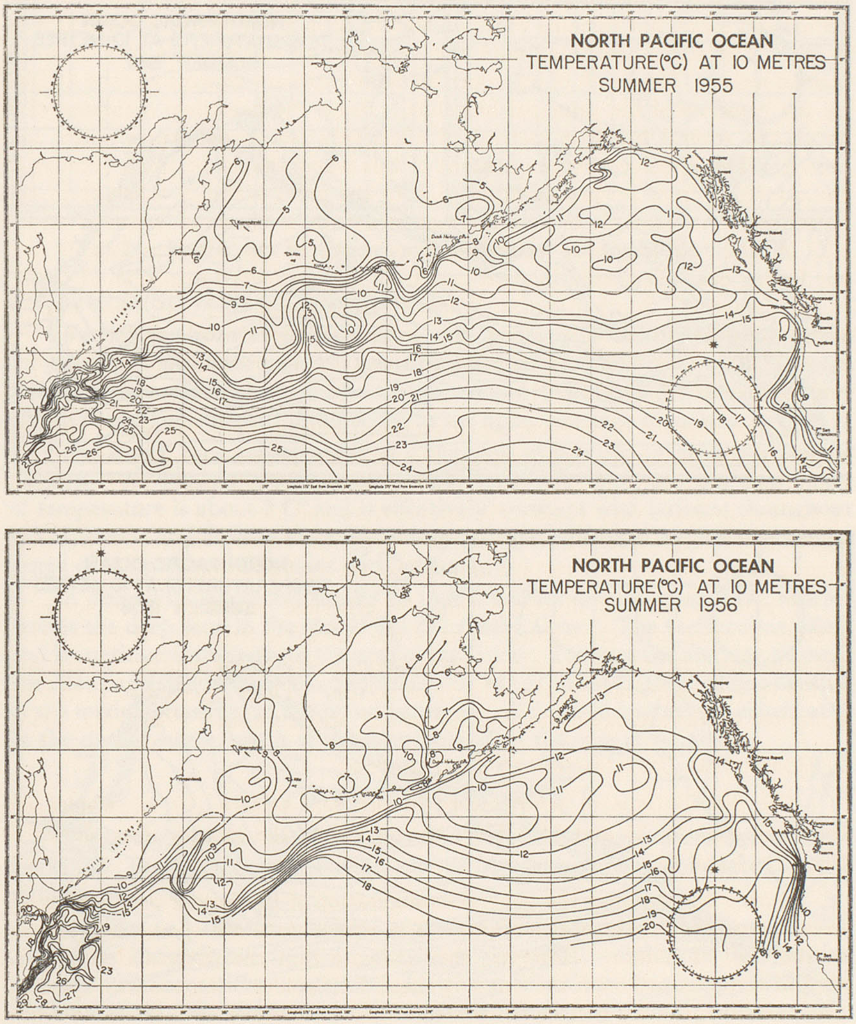


FIG. 24. Temperature at 10 metres, summer 1955 and summer 1956.

Off the northern Japanese islands there are several "filaments" of closely-spaced isotherms. These indicate the distinct fronts associated with the confluence of the warm Kuroshio and cold northern waters (page 127). Actually, the conditions there are much more confused than shown, because of the turbulent nature of the currents and the short term variability in the boundaries between the water masses (Uda, 1938a, b).

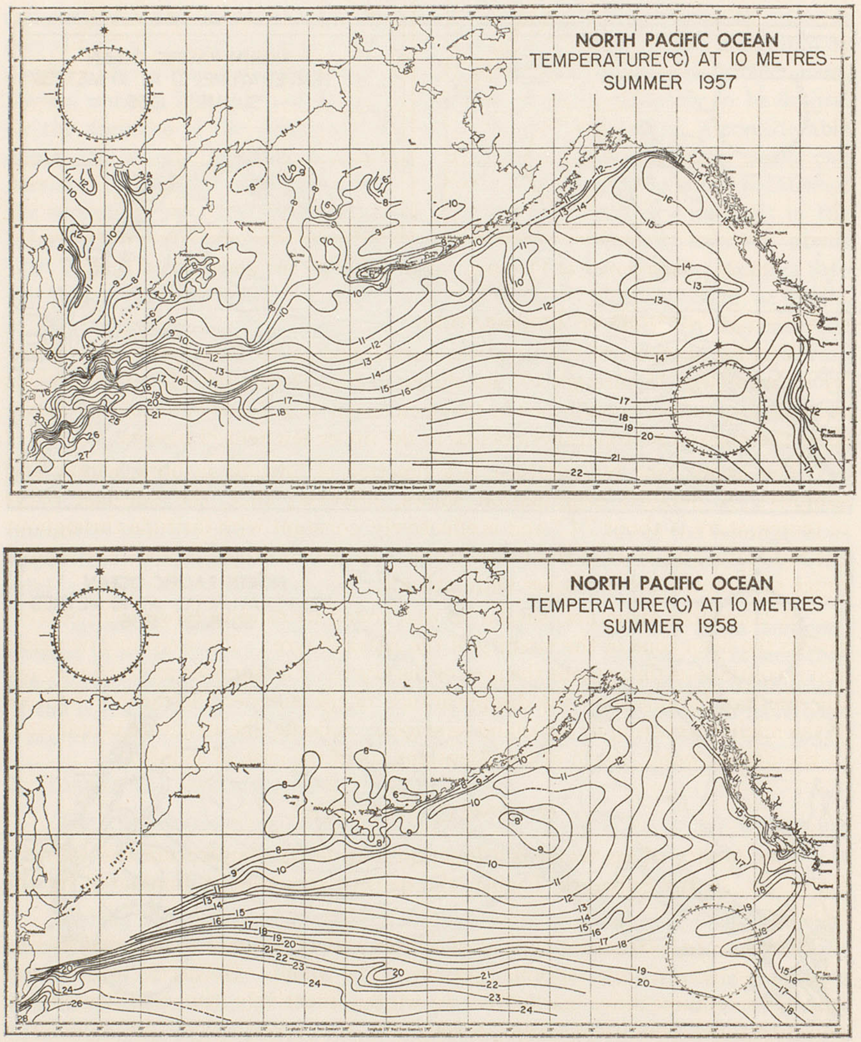


FIG. 25. Temperature at 10 metres, summer 1957 and summer 1958.

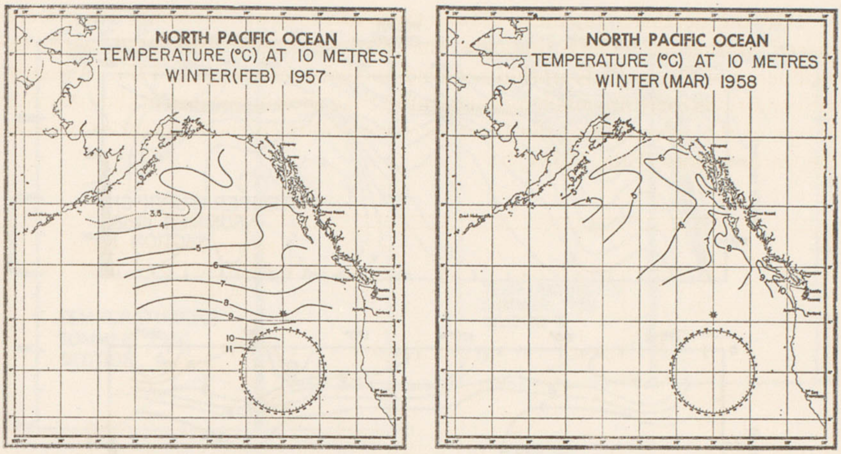


FIG. 26. Temperature at 10 metres, winter (Feb) 1957 and winter (March) 1958.

3. Vertical Sections of Temperature

Vertical sections of temperature, corresponding to those previously given for salinity (Fig. 15 through 19) are shown in Fig. 27 through 31. The sections representing summer conditions in the open ocean (e.g. Lines 8, 9 and 10) clearly show the marked seasonal thermocline in the upper few tens of metres.

The upper zone temperature is a function of latitude in both winter and summer. A detailed examination shows, however, that the seasonal *range* of temperature is about 7 C° and is effectively constant with latitude throughout the Subarctic and most of the Polar Front. In the Subtropic Region, the annual range of temperature decreases with latitude.

A dome-like structure, similar to that found for salinity (page 139), characterizes the deep zone in the vicinity of the Alaska Gyre. The thermocline exists and is continuous throughout the area of the Gyre. The summer sections through the Aleutian Islands show the disappearance of the summer thermocline because of the mixing occurring in the island passages. A dome-like structure is indicated in the deeper water, south of the Aleutians, as in the case of salinity.

G. DENSITY STRUCTURE

The density of sea water is a function both of its temperature and salinity. Generally, in the Subarctic Pacific, the density structure is dominated by the salinity.

Density (σ_t) structure typical of winter and of summer is depicted in Fig. 32. σ_t is defined by: (specific gravity -1×1000). Subarctic winter and summer structures differ markedly. In winter, effectively isopycnal (isothermal and isohaline) conditions occur throughout the upper zone. In summer a minor pycnocline, corresponding to the shallow thermocline (page 145) appears.

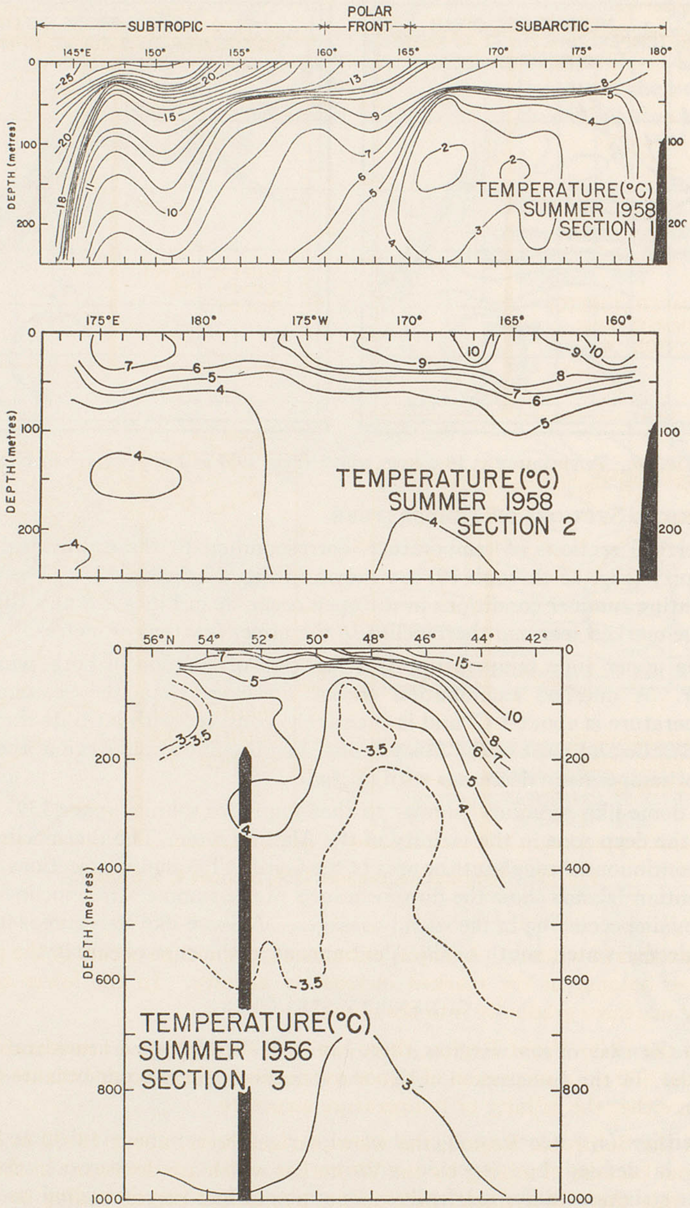


FIG. 27. Temperature, Section 1, summer 1958, Section 2, summer 1958 and Section 3, summer 1956.

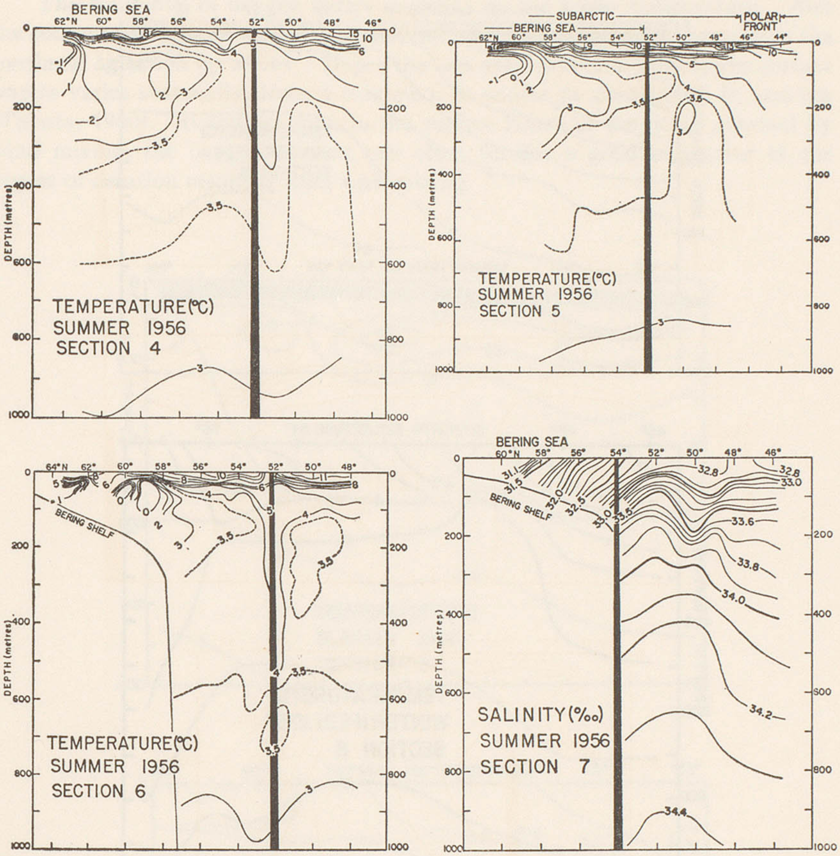


FIG. 28. Temperature, summer 1956, Sections 4, 5, 6 and 7.

This vanishes in winter. A major, permanent, pycnocline occurs in the principal halocline because of the marked increase in salinity. In the lower zone, the density increases gradually into the abyss.

In the Polar Front Region, there can exist throughout a considerable depth (to about 400 m) a succession of minor pycnoclines; these occur because of inversions in either the salinity or temperature structure.

In the Subtropic Region, the warm surface layer, about 10–30 m depth, is isopycnal because of mixing by the Westerlies which are dominant in the region. Below this there is a single, permanent, pycnocline corresponding to the permanent thermocline. In this region the density structure is temperature-dominated.

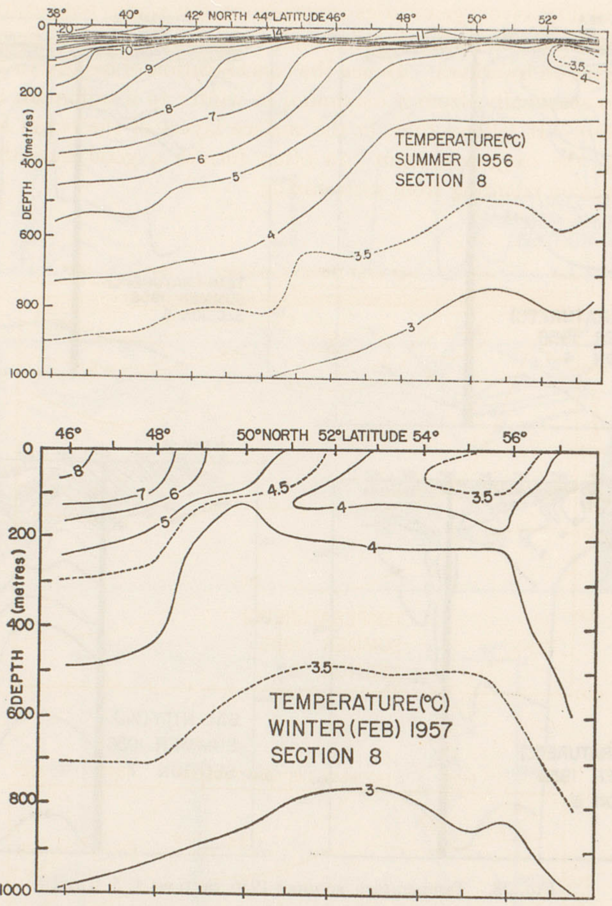


FIG. 29. Temperature, Section 8, summer 1956 and winter (Feb) 1957.

H. DISSOLVED OXYGEN CONTENT OF SUBARCTIC WATERS

The dissolved oxygen content of sea water is one of the basic indicators of marine photosynthetic and respiratory activity—of both flora and fauna—in upper layers. In the euphotic zone (0–50 m) oxygen maxima arising from floral plankton production can appear, given favourable conditions of entrant light, temperature, and dissolved nutrients. Low concentrations or minima of dissolved oxygen at depth (600–1000 m) can be attributed mainly to zooplankton (faunal) respiration and to the decomposition of sinking organic material. The oxygen content in the very deep waters in generally related to non-biological processes.

The solubility of oxygen varies inversely as the water temperature. Also the concentration in the surface waters tends to be near saturation because of the continual agitation by waves. Hence the concentration of oxygen in the surface waters varies seasonally, from a maximum in winter to a minimum in summer (Tabata, 1960). It follows that, in the surface layers of the ocean affected by wind mixing, the oxygen content can often furnish a good indication of the degree of aeration resulting from such mixing.

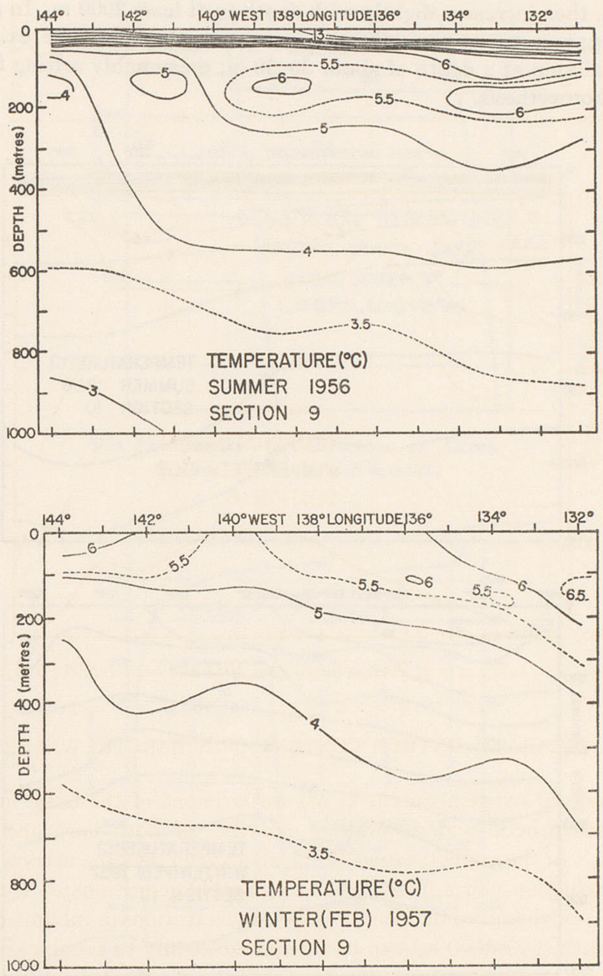


FIG. 30. Temperature, Section 9, summer 1956 and winter (Feb) 1957.

1. Structure of the Dissolved Oxygen Content

Typical winter and summer structures of dissolved oxygen content in the Subarctic Pacific are shown in Fig. 33. The large, quite uniform content (greater than about 0.6 mg-at/l) occurring in winter in about the upper 100 m arises both from the colder water temperature and from the efficient aeration by the strong wind mixing occurring at this time. This iso-oxygen layer coincides with the isohaline and isothermal layer. Below this is a sharp oxycline, extending to about 300 m depth, through which the values decrease to 0.1 mg-at/l. The concentration then usually decreases to a minimum (0.03 mg-at/l) between 800 and 1000 m, then increases slightly with depth to at least 2000 m. In summer the structure appears to be basically the same, except that there exists a fairly marked maximum at a depth of about 30–50 m, presumably arising from phytoplankton photosythesis.

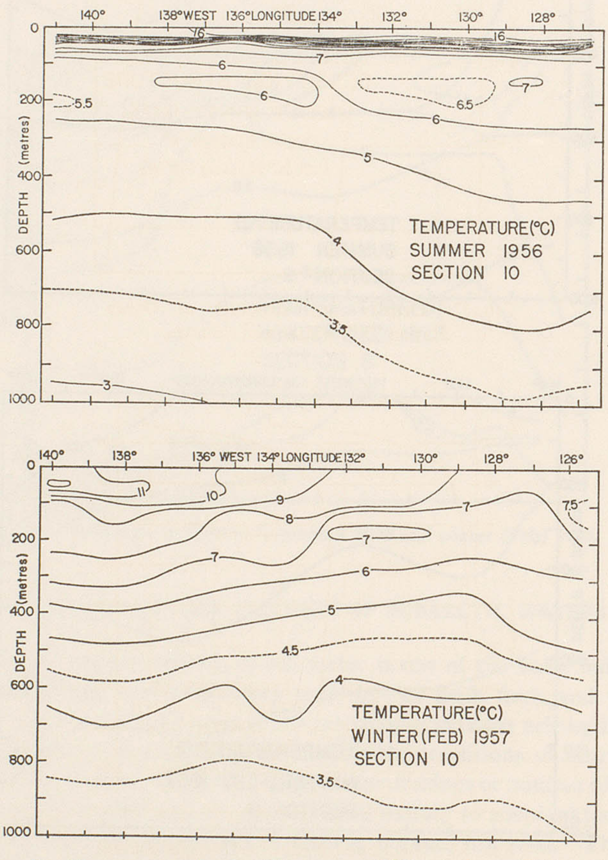


Fig. 31. Temperature, Section 10, summer 1956 and winter (Feb) 1957.

Near the western boundary, the oxygen structure in the Polar Front is often marked by a second maximum (not shown). This occurs because of the presence at depth of highly-oxygenated waters, such as those of the cold Oyashio Undercurrent. In the Subtropic Region, the shallow maximum due to phytoplankton production is rare because of the generally small productivity in this region.

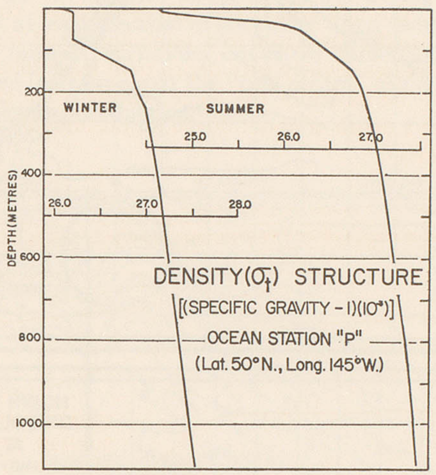


FIG. 32. Density (σ_t) structure at Ocean Station "P", winter and summer.

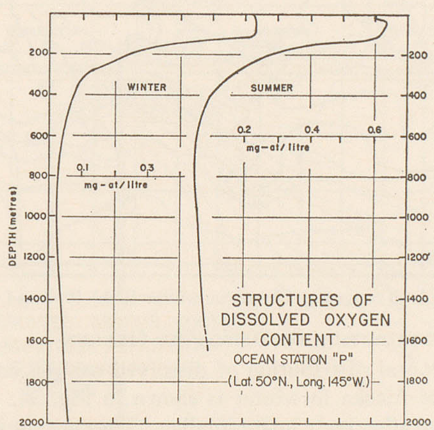


FIG. 33. Structures of dissolved oxygen content at Ocean Station "P", winter and summer.

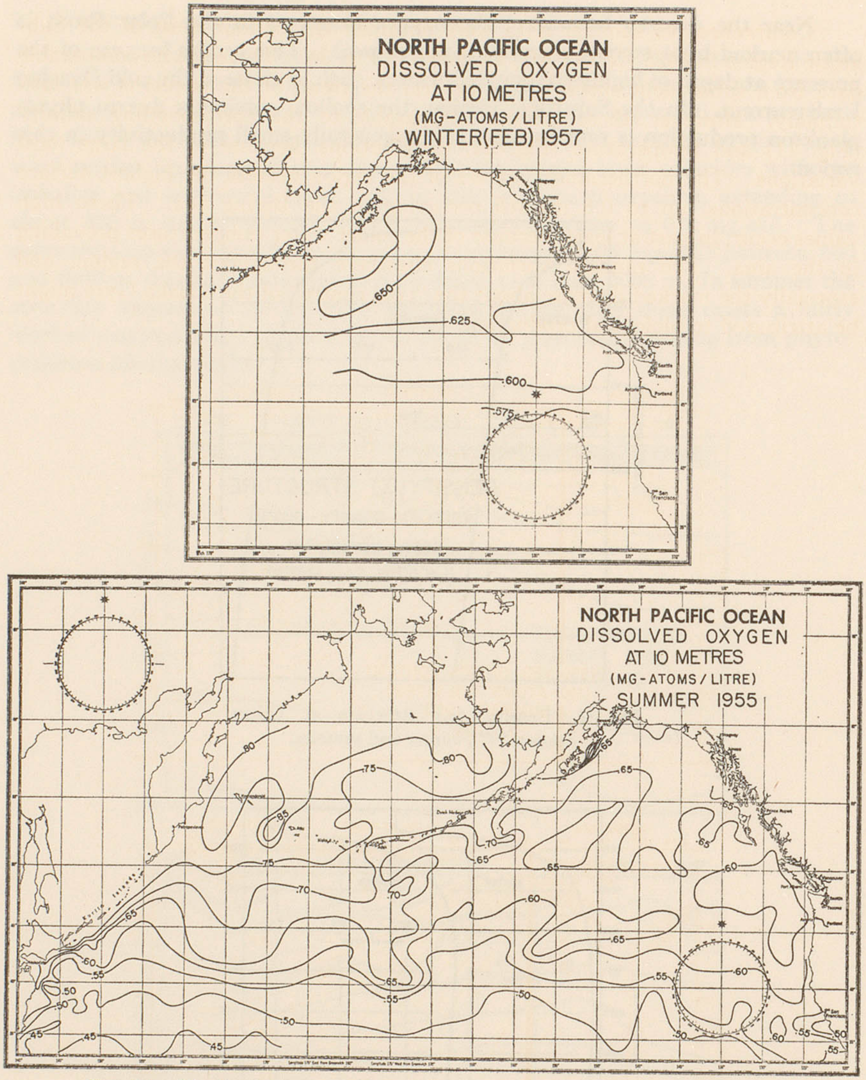


Fig. 34. Dissolved oxygen at 10 metres, winter (Feb) 1957 and summer 1955.

2. Horizontal Distribution of Dissolved Oxygen

A typical horizontal distribution of dissolved oxygen content at 10 metres depth in the Gulf of Alaska in winter is shown in Fig. 34. The data indicate that the isopleths run effectively latitudinally. The content increases northward from the Subarctic boundary to the Alaska Gyre. The highest concentration is associated with the low-temperature water in the dynamic centre of the Gyre.

Comprehensive summer data were obtained throughout most of the Subarctic Region in 1955 (NORPAC) as shown in Fig. 34. There are no comparable data from subsequent surveys. In general the trend of the isopleths was latitudinaly. However, these are not comparable to the corresponding salinity (Fig. 11) and temperature (Fig. 24) distributions, because of the large local variations. The northernmost waters of the region possess the highest surface concentrations, up to 0.8 mg-at/l occurring in the Bering Sea. These high values occur, in part at least, because of the colder water temperatures prevailing in the area. The random variations elsewhere throughout the region, and especially the marked increase about the "cold centre" of the Alaska Gyre might be attributed to biological production.

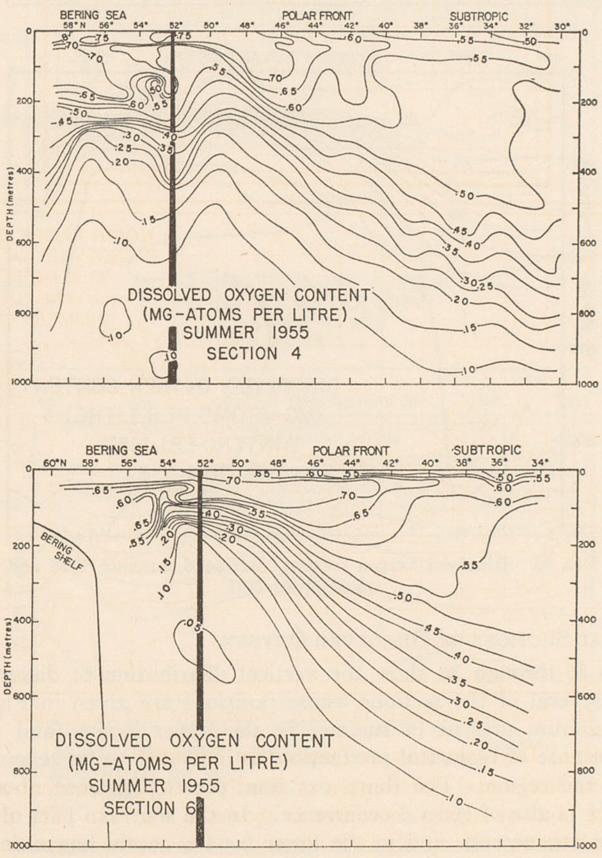


FIG. 35. Dissolved oxygen content, summer 1955, Section 4 and Section 6.

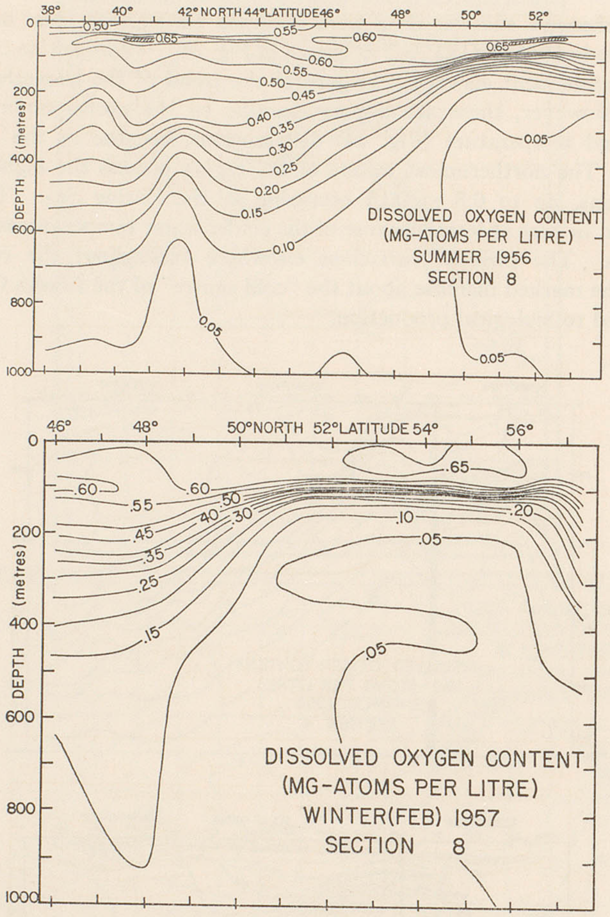


FIG. 36. Dissolved oxygen content, Section 8, summer 1956 and winter (Feb) 1957.

3. Vertical Sections of Dissolved Oxygen

Figures 35 through 38 show the vertical distribution of dissolved oxygen content in several of the sections whose positions are given in Fig. 14. The marked maximum present in summer in the upper layers (and presumably occurring because of biological production) is indicated to be generally present throughout the region. The sharp oxycline, present between about 100 and 300 m depth, is also of general occurrence. In the southern part of the region, the oxycline degenerates, and in the lower zone a nearly linear decrease with depth occurs. In the area of the Alaska Gyre, the lower-zone dome structure, a feature of the distributions of both salinity and temperature, is also present in the oxygen distribution. The oxygen minimum present in the eastern part of the region at depths between 800 and 1000 m (Section 8, Fig. 36), rises in the dome to within 400 m of the surface. The near-surface oxycline remains unchanged within the area of the dome. Sections 4 and 6 in the central portion of the Subarctic (Fig. 35) show decreasing oxygen content with depth. They also show that the isopleths of low concentration in the deep water ascend northward, similar to the isohalines (Fig. 16) and the isotherms (Fig. 28). As shown in Sections 4 and 8 (Fig. 35 and 36) the oxycline is generally present throughout both the Polar Front and Subtropic Regions. It is compressed and becomes shallower towards the north.

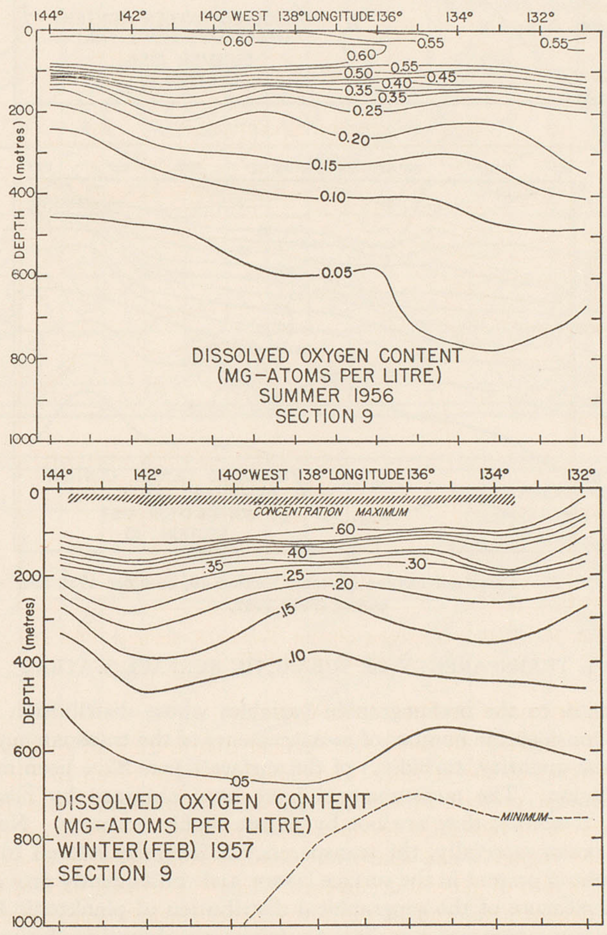


FIG. 37. Dissolved oxygen content, Section 9, summer 1956 and winter (Feb) 1957.

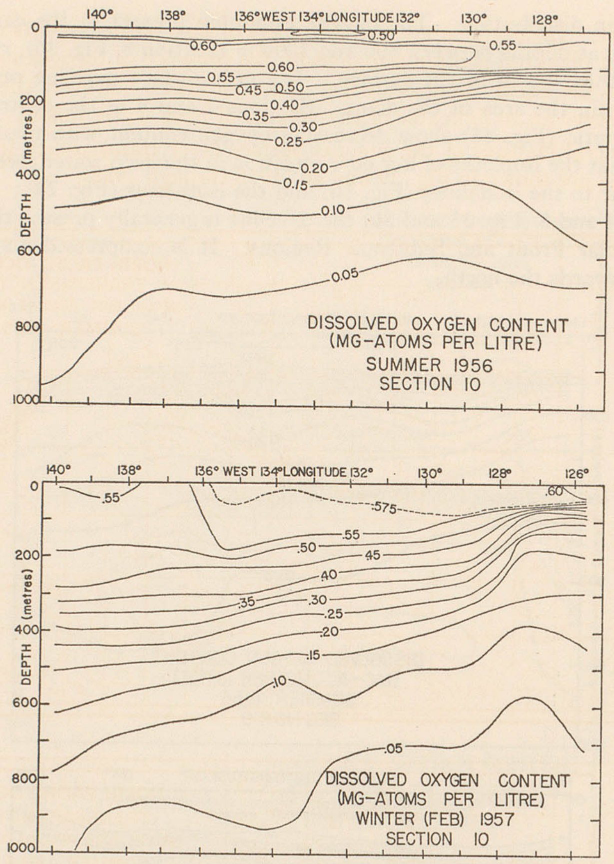


FIG. 38. Dissolved oxygen content, Section 10, summer 1956 and winter (Feb) 1957.

I. TRANSPARENCY OF SUBARCTIC SURFACE WATERS

In addition to the oceanographic variables whose distribution have been discussed, a considerable number of measurements of the transparency or clarity (or the inverse quantity, turbidity) of the surface layers have been made in the Subarctic Region. The measurements have been obtained by means of the Secchi Disc; therefore, they are not by nature highly accurate. Nevertheless, in the open ocean especially, the transparency is inversely related to the abundance of plankton present in the surface layers, and consequently provide a rough quantitative measure of the geographical distribution of planktonic life. Thus transparency measurements are often useful in corroborating the productivity indicated by the relative values of the dissolved oxygen content found in the surface layers. In coastal areas, the usefulness of such measurements to indicate abundance of plankton is often impaired by the presence of terrigenous material brought into the sea by land drainage. In such areas, however, the measurements may serve as an indicator of the volume of fresh water inflow.

The general features of the summer surface layer transparency in the Subarctic Region are shown in Fig. 39 and 40 (1955 through 1958).

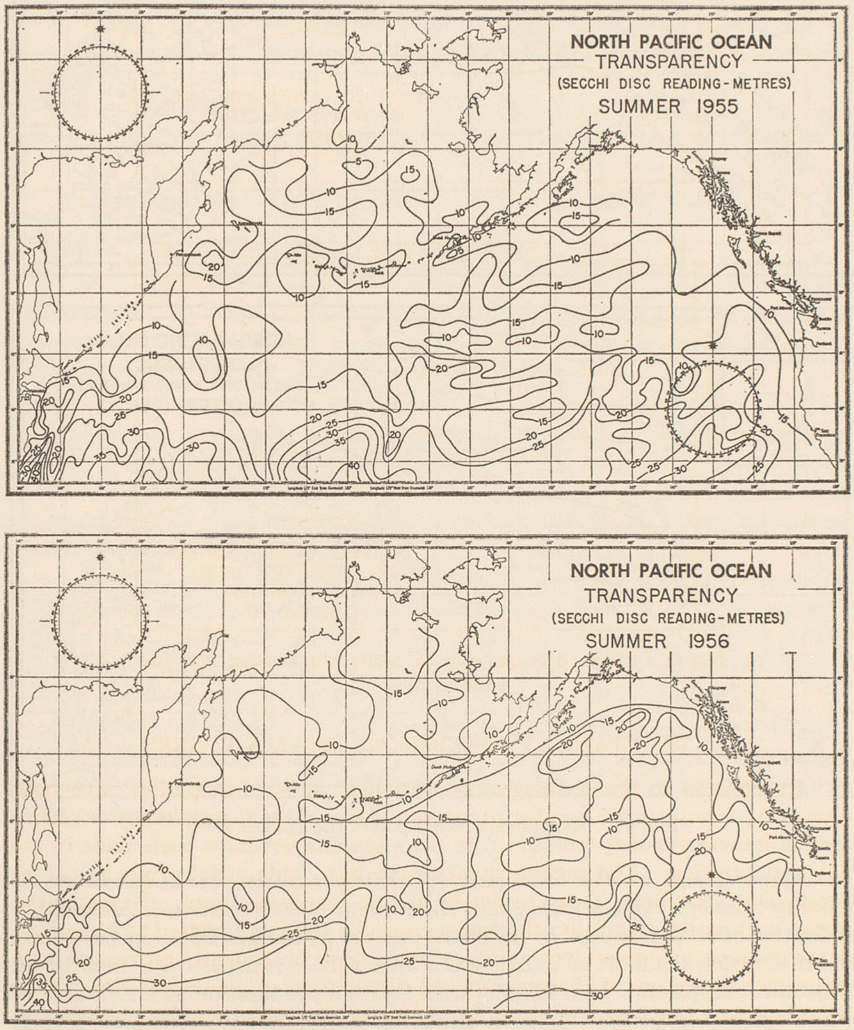


FIG. 39. Transparency, summer 1955 and summer 1956.

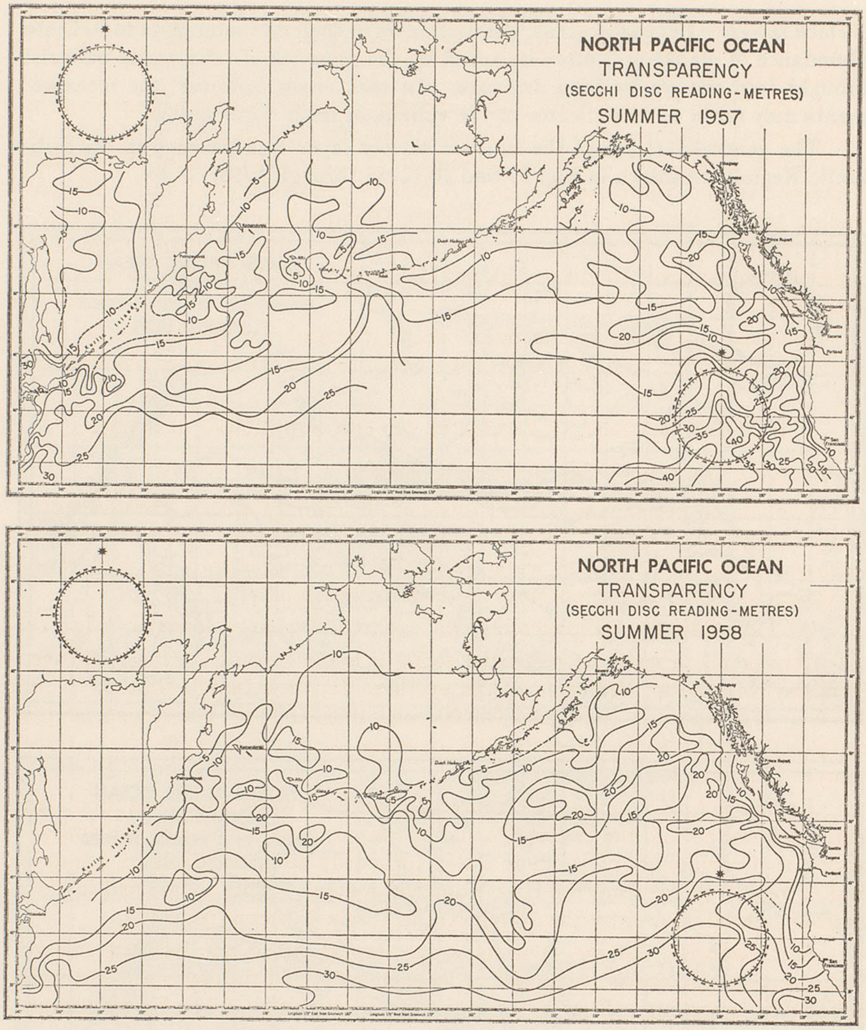


FIG. 40. Transparency, summer 1957 and summer 1958.

Marked, and for the most part apparently random, variations of transparency are very evident throughout the region. This condition is presumably indicative of a "patchiness" in the distribution of planktonic life. In the waters of the Oyashio the limit of visibility of a Secchi Disc is usually less than 10 m. This low transparency is in keeping with Oyashio's reputation as a rich area of planktonic life (page 167). The low-salinity coastal waters are even more turbid, visibility being generally much less than 10 m. This fact is due, in some measure at least, to silt. Localities of relatively high transparency (15–20 m) approximate in position the central high-salinity cores both of the Bering Sea and of the Alaska Gyre.

In the Polar Front Region, there appears a marked southward increase in transparency, visibilities to 30 m being common (and some as great as 45 m being obtained occasionally). These values indicate that a smaller amount of plankton is present in the Subtropic Region. A belt of relatively clear water (Secchi reading greater than 20 m) present between Ocean Weather Station "P" (Lat. 50° N, Long. 145° W) and the American coast, provided further evidence of unusual oceanographic conditions (page 171) in 1957 and 1958 in that area (Tully *et al*, 1960).

Generally speaking, in the open ocean, transparency appears to be proportional to salinity, and inversely proportional to concentration of plankton.

J. OTHER IMPORTANT FEATURES OF THE SUBARCTIC REGION

1. Major Currents of the Subarctic Region

Two major permanent current systems lie entirely within the Subarctic Region (page 120) (Fig. 7). Both currents are affected to some degree by seasonal (non-persistent) intrusions of water from the Polar Front into the Subarctic. Characteristics of the two currents are summarized briefly:

(i) The Oyashio (Japanese: "Oya"—parent, "shio"—current) (Fig. 7).

Considered in its most restricted sense, it is the current flowing southwest from the Kamchatka Peninsula to the northern Japanese islands (Lat. 50° N, to 40° N). Transport is between 7,000,000 and 10,000,000 m^3 per second, and is composed, at the southern extremity, of equal amounts of water from three sources (Sugiura, 1957):

East Kamchatka Current—This current arises from water which has circulated anti-clockwise around the Bering Sea and has been greatly affected by the appearance (in the summer) of ice-melt water (primarily from the Gulf of Anadyr) and of land drainage from the East Kamchatka coast. This is modified in turn both by upwelling and by the North Kurile Warm Current (page 171).

East Sakhalin Current—This is water flowing along the western side of Okhotsk Sea. It too is composed of ice-melt and land drainage water and is affected by upwelling. Flow of this water out through the middle Kurile Islands continues throughout the year (page 150).

Oyashio Gyre—Water recirculated in this cyclonic gyre is present off the Kurile Islands between about 150 and 155° east longitude.

Major Characteristics of Oyashio. The temperature is low, generally less than 9° C. Salinity is low, less than 33‰. The depth of the current is 200–400 m. The speed is between 0.3 and 1 knot. The nutrient content is high. The Japanese name may originate from the fact that the current is a rich source of food for marine life. The transparency is low; the Secchi Disc visibility is usually less than 15 m.

A major portion of the Oyashio turns east and forms part of the West Wind Drift. A minor part forms the Oyashio Undercurrent which moves south and southwest under the Kuroshio. It is believed that the strength of the Undercurrent is determined primarily by meteorological conditions (page 173).

(ii) The Alaska Stream and Its Extension.

The Alaska Stream is the narrow intense stream flowing along the southern coast of the Alaska Peninsula from Kodiak Island to Unimak Pass (Long. 150° W to 160° W). Its speed is of the order of one-half knot. The transport includes most of the flow in the Alaska Gyre plus the northern transport adjacent to the Canadian Coast. In summer, the total transport is about 17×10^6 m^3 per second and flow occurs to at least a depth of 2000 m (Bennett, 1959). Intensification characterizing the stream is probably due to the conservation of absolute vorticity through changing latitude. It appears, from results of drift-bottle releases (Dodimead and Hollister, 1958), that a portion of the Alaska Stream recirculates around the Alaska Gyre at least occasionally, although this is not readily apparent in the geostrophic circulation (Fig. 5 and 6). The Alaska Stream Extension is the continuation of the Stream westward past Unimak, and may be termed the *Return Flow* (of the West Wind Drift) out of the Gulf of Alaska. The Extension develops westward in the spring and summer, reaching furthest west about August. The development is due to the increase of land drainage from the southern coast of Alaska during the spring and summer months. Water moves out to the end of the Aleutian Chain, successive portions moving through the island passes into the Bering Sea. Extension water moving through the Aleutian passes is affected to a degree by the (seasonal) South Aleutian Warm Current.

Major characteristics of the Alaska Stream. The temperature of both the current and its extension is relatively warm throughout the year. The salinity is low throughout (especially in summer because of land drainage) but increases along the line of flow due to entrainment (page 169) and mixing.

2. Cold Cores

In the Subarctic Region, there are several localized areas in which the upper-zone water exhibits temperatures markedly less than those of the surrounding water, during part or all of the year. These areas are depicted in Fig. 41 and are termed "Cold Cores". They are assumed to be formed by a number of factors acting either singly or together, such as: southward intrusions of cold water into an area of warmer water, the formation of ice-melt water, upwelling, and the movement of deep zone water into the upper layers.

The Core numbered (1) appears to be a manifestation of general southward intrusions of branches of the cold Oyashio Current. One such intrusion, presumably generated by meteorological conditions, is discussed more fully on page 173. Cores (3), (4), and (5) are formed by ice-melt water released in the summer. The tendency to lower temperatures is also enhanced by the upwelling associated with the south and southwest winds prevailing at this time. Such upwelling

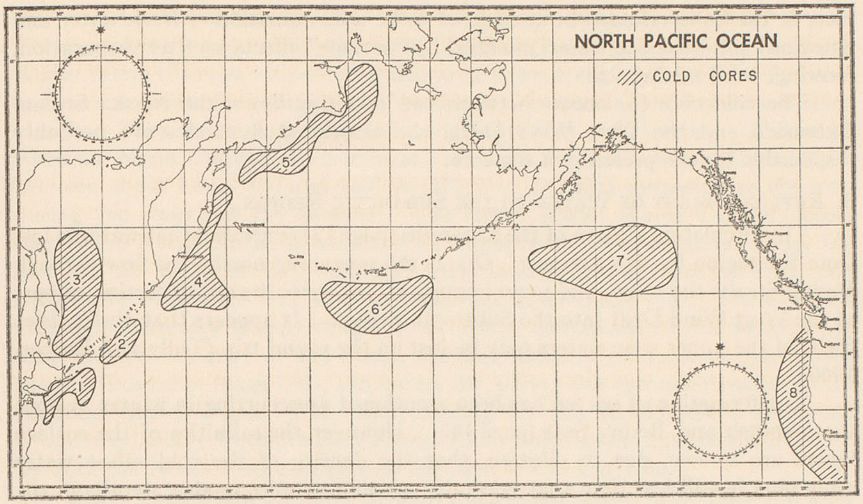


FIG. 41. Cold cores.

draws water to the surface from modest depths only, usually less than 200 m. Core (8) off the Oregon and California coasts between Lat. 30 and 45° N occurs because of the prevailing northwest winds of spring and early summer, which cause upwelling during the period from March to July.

Cores (2) and (7) occur near the dynamic centre of the Oyashio and Alaska Gyres respectively. In the Alaska Gyre core, the one most intensively studied, several important features have been noted. The isolines of all properties in the deep water exhibit at all times pronounced domelike configurations (page 162). However, the typical subarctic structure (Fig. 8) is present and continuous in the dome. Both the principal halocline and the thermocline exist there. The depth of the upper zone is slightly less than that in the surrounding area (Dodimead, 1961).

The typical Subarctic structure is similar to that present in the coastal inlets of British Columbia. In these inlets, velocity differences between upper and lower zones gives rise to a shear between them. This shear apparently provides the energy for the preferential upward transfer or *entrainment* of deep water into the upper zone (Tully, 1958). In this concept, the shear zone is in the transition zone (the principal halocline).

It appears, both from the continuation of the typical Subarctic structure through the area of the Alaska Gyre and from the horizontal distribution of properties in the area, that sub-halocline water is entrained into the upper zone in a manner essentially similar to that in British Columbia inlets (Tully and Barber, 1960). Tully (private communication) contends that the most intense entrainment appears to occur in the dynamic centre of the Gyre. Presumably, a localized centre of entrainment is also present in the Oyashio Gyre. The upper zone in the Subarctic can be considered to be a combination of fresh water and entrained sub-halocline water, modified by "surface" effects such as evaporation, freezing, wind mixing, etc.

The cold core (6) occurs between two opposing flows, the Alaska Stream Extension and the West Wind Drift; the associated dynamics are probably responsible for the presence of the core.

3. Replenishment of Water to the Subarctic Region

The circulation system of the Subarctic (page 127) indicates that water is lost from the region by two avenues. One is the prevailing northward flow through Bering Strait; the other, and more important, is the southward deflection of part of the West Wind Drift into the Subtropic Region. It appears that about three fifths of the upper zone waters may be lost on the round trip (Tully and Barber, 1960).

The formation of sea ice has been mentioned as occurring in winter in both the Okhotsk and Bering Seas (page 147). However, the salinities of the surface layers are so low, due to dilution, that the density of the cold saline water remaining is still less than that generally present at quite modest depths in the Subarctic Region. Deep water is therefore not formed by convective overturn (and indeed, is not produced at all) in the Subarctic Pacific.

Masuzawa (1960) studied the Oyashio Undercurrent off the coast of Japan and concluded that in this limited region (Cold Core (1), Fig. 41) the Oyashio water descended to about 1100 m depth. Below this, the abyssal waters were generally rising. However, this observation bears on the genesis of intermediate waters in the Polar Front and Subtropic Region, which has not been discussed in this review. It in no way detracts from the conclusion that deep water is not produced in the Subarctic Region as defined here (Fig. 1). The region thus differs from others such as the North Atlantic and the Antarctic, in which large quantities of deep and bottom water originate at the surface. Any renewal, therefore, of deep water which has been entrained or upwelled into the upper zone and subsequently lost from the Subarctic Region because of the persistent circulation, can only occur by deep inflow from south of the region, and by subsequent upwelling or entrainment into the upper zone.

In his recent study, Bennett (1959) indicated that at least one source of persistent deep flow into the Subarctic exists. It appears that below the halocline (200 m) down to at least 2000 m depth, water persistently intrudes into the Gulf of Alaska from the southeast and circulates around the deep basin.

4. Intrusions in the Subarctic Region

Several non-persistent flows of water occur in the upper few hundred metres, and influence the persistent circulation quite markedly. In addition, non-persistent movements are presumably present in the deep and abyssal waters; however, nothing is known about their characteristics up to this writing. It is theorized, however (page 174), that the movements in the upper layers originate primarily from the meteorological conditions that occur in the Subarctic Region and its surroundings.

Two (warm) intrusions from the south occur during the summer period. These weak surface currents, the South Aleutian Warm Current and the North Kurile Warm Current, are probably due to summer strengthening of the Kuroshio and the coincident weakening of the Oyashio. Commencing in May they develop rapidly. They decay during the following September and October. The South Aleutian Warm Current is a northward intrusion from about Lat. 45 to 50° N, between about Long. 170 and 180° W (Fig. 7). It shifts somewhat to the west during the course of the summer. The North Kurile Warm Current occurs east of the North Kurile Islands and off Kamchatka, between about Long. 160 and 165° E. Some of the (relatively) warm saline water of these currents cools sufficiently to subside and form mesothermal water (page 150). The currents, although transitory in nature, affect the overall circulation of the Subarctic to some degree. A part of the South Aleutian Current mixes with the Alaska Stream Extension (page 168), and enters the Bering Sea through passages in the Aleutian Island Chain. The water entering the Sea of Okhotsk is composed of water originally from the East Kamchatka Cold Current and thus is modified by both the North Kurile and South Aleutian Warm Currents.

While these warm currents recur to some extent every year, their intensity appears to vary markedly with the years. For example, summer surface water temperatures in the Western Bering Sea were found during the period 1954–58 to be warmest in 1957 and colder in 1955, 1958, and 1956, respectively (Fig. 14 and 25). The variation in these temperatures is presumably due primarily to fluctuations in the intensity of the North Kurile Warm Current.

Other northward intrusions, whose intensity appears to vary more slowly with time than those just mentioned, occur at both eastern and western sides of the Subarctic Region. The eastern, the more massive one, occurs between about Long. 150° W and the North American Coast. The western one occurs off the northeastern coast of Japan at about Long. 145° E.

The North American Intrusion has been studied by Tully *et al.* (1960). In 1955 and 1956, the division of the West Wind Drift occurred off the American Coast between Lat. 46 and 48° N, well within the Subarctic Region. In 1957 it was noted that the division had shifted, since the beginning of the year, about 200 miles southward. Thus more of the relatively warm (Polar Front) water entered the circulation of the Alaska Gyre. The geopotential topography indicated that the mid-ocean currents tended more northward than in previous years. This alteration in flow was confirmed by drift bottle releases. At the same time, transparency measurements indicated that, on the surface at least, very clear water indicative of Polar Front water appeared in the general area of the intrusion.

The warming was found to extend to about 500 m depth. This fact signifies that the phenomenon resulted principally from the influence of advection and not from seasonal heating and cooling effects, which are not significant much below a depth of about 100 m (Tully *et al.*, 1960). Fig. 42 shows the variation in the extent of the intrusion during six successive surveys in the eastern Subarctic throughout the period 1955–58. In this Figure, the temperature distribution on the surface of constant density σ_t = 26.60 was examined. This surface is

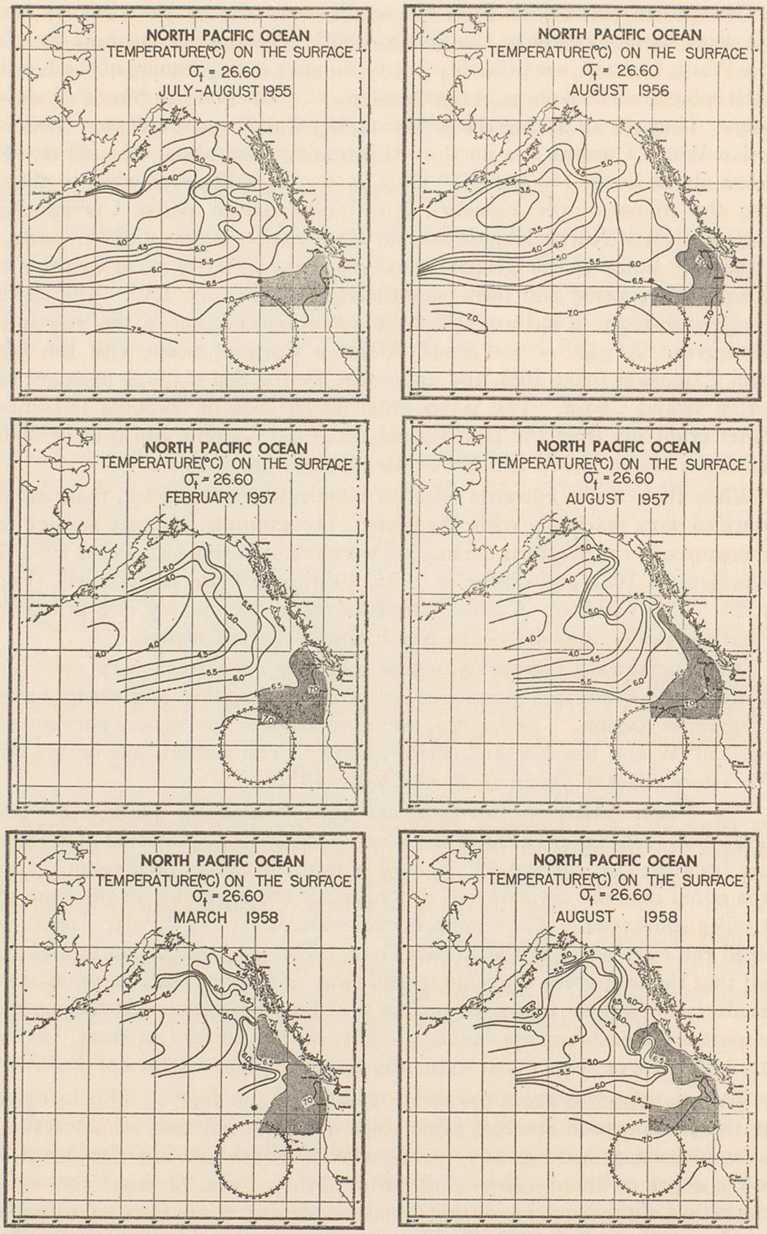


FIG. 42. Temperature on the surface of σ_t = 26.60, July–August 1955, August 1956, February 1957, August 1957, March 1958 and August 1958.

known to lie between 125 and 225 m depth in the area of interest. Thus it is always below the depth of seasonal influence. The shaded portions show the progressive increase of the area of water warmer than 6.5°C. The relative strength of the intrusion was, by years, 1958 (strongest), 1957, 1955 and 1956 (weakest).

The intrusion of warm water northward along the Japanese coast was reported by Uda (1938a, 1960) and Takenouti (1960). By contrast with the sequence off the Canadian coast, it was 1955 (strongest), 1956, 1957, 1958 (weakest). In addition, the weakening of this intrusion appeared to be accompanied by the simultaneous development of a countering southward movement of colder water. This occurrence was caused by variation in strength of the southward-flowing Oyashio Current and Undercurrent.

A fact readily noticeable is the phase difference of occurrence of the various intrusions. There appears to be a lag of about 3 years between the extremes of the intrusions in the western and eastern Subarctic respectively.

A relatively simple explanation, one which appears to relate firmly the strength of the intrusions in the Subarctic and the meteorological conditions in the region and its surroundings, can be put forward. The explanation avails itself of the fact that wind speed (W) is virtually proportional to the square root of the difference between "core-pressures" of opposing barometric high- and low-pressure cells in the atmosphere. Therefore,

$$W = k_1 \sqrt{P_H - P_L} \qquad \ldots (1)$$

P_H = core-pressure of high-pressure cell

P_L = core-pressure of low-pressure cell

The speed (V) of the current generated by the resulting wind stress is in turn proportional to the wind speed,

$$V = kk_1 \sqrt{P_H - P_L} = KW \qquad \ldots (2)$$

where K is the "wind factor", having a value between 0.02 and 0.03. The wind-induced current in the sea is, to a first approximation, parallel to the isobars of atmospheric pressure. (The direction of the wind at sea level is about 30° to the left of the isobars, while the wind driven current is about 30° to the right of the wind), (Uda, 1955).

In Fig. 43 are shown average monthly differences in core-pressure between the various cells influencing meteorological conditions in the Subarctic Region (page 122, Fig. 2 and 3). The period covered is from 1955 to 1958 inclusive; the quantities plotted should be proportional to the square of the resulting wind speed. Fig. 43(1) shows the core-pressure difference between the Siberian High and the Aleutian Low; the resulting winds must tend to affect conditions on the western side of the Subarctic Region. The area under the "Pressure-difference versus Time" curves is taken to be indicative of the transport directly induced by wind. The transport appeared in this case to be southward most of the year, the effect being especially marked during the winter months. The annual transport increased throughout the 4-year period, reaching a maximum in the winter of 1957–58. This fact is presumed to account for the progressively

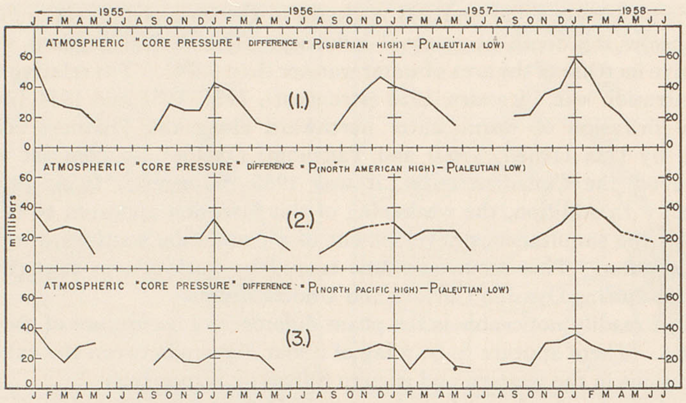


FIG. 43. Atmospheric "core pressure" differences.

colder water temperatures encountered, in succeeding summers, off the northeast coast of Japan. (It also gives a measure, therefore, of the weakening of the warm intrusion.)

Figure 43(2) shows core-pressure differences between the North American High and the Aleutian Low, and Fig. 43(3) shows those between the North Pacific High and the Aleutian Low. The differences in Fig. 43(2) which would give rise to effectively northward flow in the eastern Subarctic, were at a maximum in winter or early spring. The effect was progressively greater through the 4-year period, and especially so during the last three of the years. The differences shown in 43(2) also lead to progressively greater northward transport into the eastern Subarctic, especially during spring. Therefore, presumably because of the combination of these two effects, the temperatures in the upper layers of the eastern Subarctic reached a maximum in 1958. In addition, the *relative* variation in transport during the 4-year period was much greater in the eastern than in the western Subarctic. This fact could explain the much more marked warming occurring in the eastern area.

The warm (northward) intrusions occurring annually in the central part of the region (the South Aleutian and the North Kurile Warm Currents) can be explained reasonably on the basis of transport induced by the pressure difference between the North Pacific High and the Aleutian Low (or, in the absence of the latter cell, by the pressure distribution in the North Pacific High itself). Such transport is present during the spring and early summer (page 00). The correlation between the annual variations in the pressure differences (Fig. 43(3)) and in the strength of the intrusions is not as marked as in the cases previously discussed. However, a pronounced minimum in the northern extent of the intrusion of 1956 is matched by a corresponding minimum in the pressure differences.

Recently the author has studied the non-seasonal fluctuation of sea-surface temperature, along the zone between 35 and 55° N latitude, across the Pacific

Ocean (Uda, 1960). This was based on the data contained in the publication "Mean atmospheric pressures, cloudiness, and sea surface temperature of the North Pacific Ocean and neighbouring seas (1911–1941)" (Imperial Marine Observatory, Kobe). He concluded that the sequence of occurrence of warmer and cooler-than-normal surface waters on the western side of the Pacific was opposite in phase to the occurrences on the eastern side.

On the western side, the minima of surface seawater temperatures occurred in 1913, 1923, 1926, 1934 and 1941. These coincided with maxima on the eastern side of the Pacific. On the western side, maxima of surface seawater temperatures occurred in 1916, 1930, 1933, and 1937. These corresponded to minima on the eastern side. The anomalies alternate from west to east across the ocean.

This converse relation between the occurrence at the western and eastern sides of the ocean of warm and cool conditions is evident in the zone between 40 and 50° N latitude, and is most obvious during the months of February, May and August. The sequence of warm and cool years is similar in the eastern Pacific and eastern Bering Sea from 40 to 55° N latitude.

Superimposed on these variations during the period 1911 to 1941 there was a general warming of 1 to 4 C° in the North Pacific waters. This was especially notable in the western and eastern sides during winter and spring. In summer the warming was most notable in the central part of the ocean, particularly in the Polar Front.

This alternating behaviour is attributed to fluctuations in the relations between the three atmospheric pressure centres (Fig. 43). In winter the Oyashio is accelerated by the monsoons generated by the pressure gradient between the Siberian High and the Aleutian Low. This cold outflow from the Okhotsk and Bering Seas becomes dominant when the pressure difference is extreme. This condition retards the northward progress of the warm Kuroshio waters and diverts them eastward into the trans-Pacific drift. On the eastern side of the ocean, the northward transport into the Gulf of Alaska depends on the pressure difference between the Aleutian Low and the North American High.

The author proposes a "pulsation" theory such that these two major gradients are intensified periodically. Hence the wind stresses are intensified and in consequence the system of gyres, which constitutes the Subarctic circulation, is intensified. This mechanism causes the intrusions. The sequence of variations is not due to a planetary wave system. Rather it is due to a quasi-stationary pulsational wave with fluctuating amplitude due to topographically confined or boundary-conditioned gyres.

Professor J. Bjerknes (1960) recently presented an interesting paper based on data observed during 1890 through 1933. He shows conditions in the North Atlantic Ocean similar to those we have observed in the North Pacific. Previously, Rodewald (1957) showed that the trends of warming and cooling occurred concurrently along the western sides of the North Pacific and North Atlantic Oceans.

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