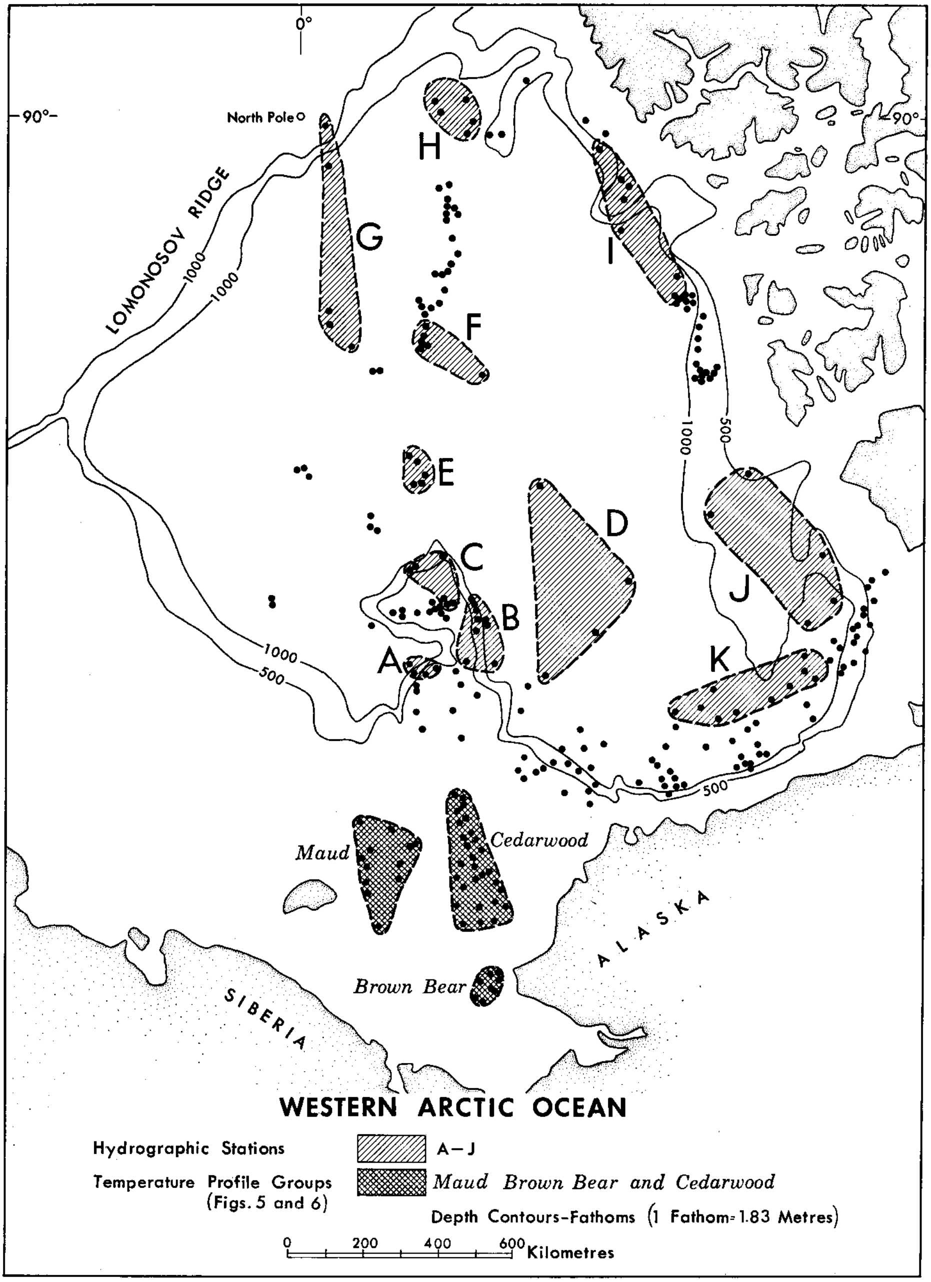

Fig. 1. Oceanographic stations and station groupings in the western Arctic Ocean.

THE CONTRIBUTION OF BERING SEA WATER TO THE ARCTIC OCEAN*

L. K. Coachman and C. A. Barnes†

WATER from the Bering Sea flows north through the narrow (74 km.) and shallow (45 metres) Bering Strait into the Arctic Ocean. The earliest measurements of this flow were made in the summers of 1932, 1933 (Ratmanov 1937) and 1934 (Barnes and Thompson 1938) and were later extended to include the winter season by Maksimov (1945). Measurements in recent years (Bloom 1956, Fleming 1959) confirm the general pattern of flow through Bering Strait established by the early workers.

According to these measurements, the transport of Bering Sea water to the north is about 1.4 million cubic metres per second in summer, and about one-fourth to one-third of this in winter. Thus the yearly average input to the Polar Basin may be estimated at 1 x 10^6 m.^{3}/sec., which is in agreement with a recent Russian estimate of 37,500 km.^{3}/year (Treshnikov 1959a). This amount of water is ten times that introduced annually into the Arctic Ocean by all large Siberian rivers (Antonev 1957, Treshnikov 1959a). If this water entered the Arctic Ocean in a layer 100 metres thick without mixing, it would occupy a strip 100 miles wide from the Chukchi Sea to the North Pole.

What happens to this water in the Polar Basin? The Russians have stated that Pacific water may be traced to the North Pole, but no substantiating evidence has been offered (Treshnikov 1959b). In what manner does the Bering Sea water enter the Arctic Ocean and how may it be traced? What role does it play in modifying the surface water, deeper water, and ice cover?

We have examined about 200 deep-water (off-shelf) oceanographic stations located in the western part of the Arctic Ocean (Fig. 1 and Table 1). For the purposes of this study the western Arctic Ocean is taken to comprise the two basins, Canadian and central (LaFond 1960), farthest removed from the Atlantic. This area is bounded by the Lomonosov Ridge,

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the Canadian Arctic Archipelago, Alaska, and Siberia. There is a concentration of stations in the area immediately north of Alaska because of extensive sampling from icebreakers. The rest of the stations were occupied from drifting ice and lie primarily around the perimeter of the Beaufort Sea. There is a paucity of observations from the interior of the Beaufort Sea and from the sector north of Siberia; only stations made from aircraft are available in these areas: the "Ski Jump" stations in the interior of the Beaufort Sea, and those made by the 1941 Russian expedition to the "pole of inaccessibility" north of Siberia.

Even though these data cover many years and all seasons of the year, there is a remarkable regularity in the vertical distribution of temperature and salinity. [It should be noted that at the rather uniformly low temperatures of the water in the Arctic Ocean the distribution of density so closely parallels that of salinity, that the latter can normally be used as an index of the mass distribution.] The variations that are observed from station to station apparently depend to a greater degree on geographic location within the basin than on secular variations. This implies that the Arctic Ocean is dynamically in a steady state and that the observed distribution of properties is a result of continuing processes within the basin.

Table 1. Deep-water oceanographic stations used in this analysis.

Station or vessel	*Number of stations*	*Source*
Fram (1894)	1	Nansen 1902
North Pole-1 (1938)	1	Shirshov 1944
SSSR-N-169 (1941)	9	A.N.-I.I. 1946
North Pole-2 (1950-1)	16	Somov 1954-55
Burton Island (1950)	35	U.S.N.H.O. 1954
Burton Island (1951)	45	U.S.N.H.O. 1954
Ski Jump (1952-3)	8	Worthington 1953
T-3 (1952-5)	11	Worthington 1959
North Pole-3 (1954)	2	Treshnikov and Tolstikov 1956
Ice Skate Alpha (1957-8)	8	Farlow 1958
Ice Skate Bravo (1957-8)	6	Farlow 1958
Ice Skate Alpha (1958)	29	English
Ice Skate Bravo (1958)	21	Collin 1959
Ice Skate Alpha-2 (1959)	23	Gast 1960
T-3 (1959)	7	Kusunoki 1959
T-3 (1959-60)	8	Muguruma 1960

Fig. 2 presents curves of temperature and salinity to show the vertical distribution of these properties at various localities in the Polar Basin. In general, the Arctic Basin contains three water masses:

(1) The surface layer (Arctic Water) has varying characteristics, but is generally cold (at or near the freezing point); it is relatively dilute at the surface but below about 50 metres the salinity increases sharply with depth.

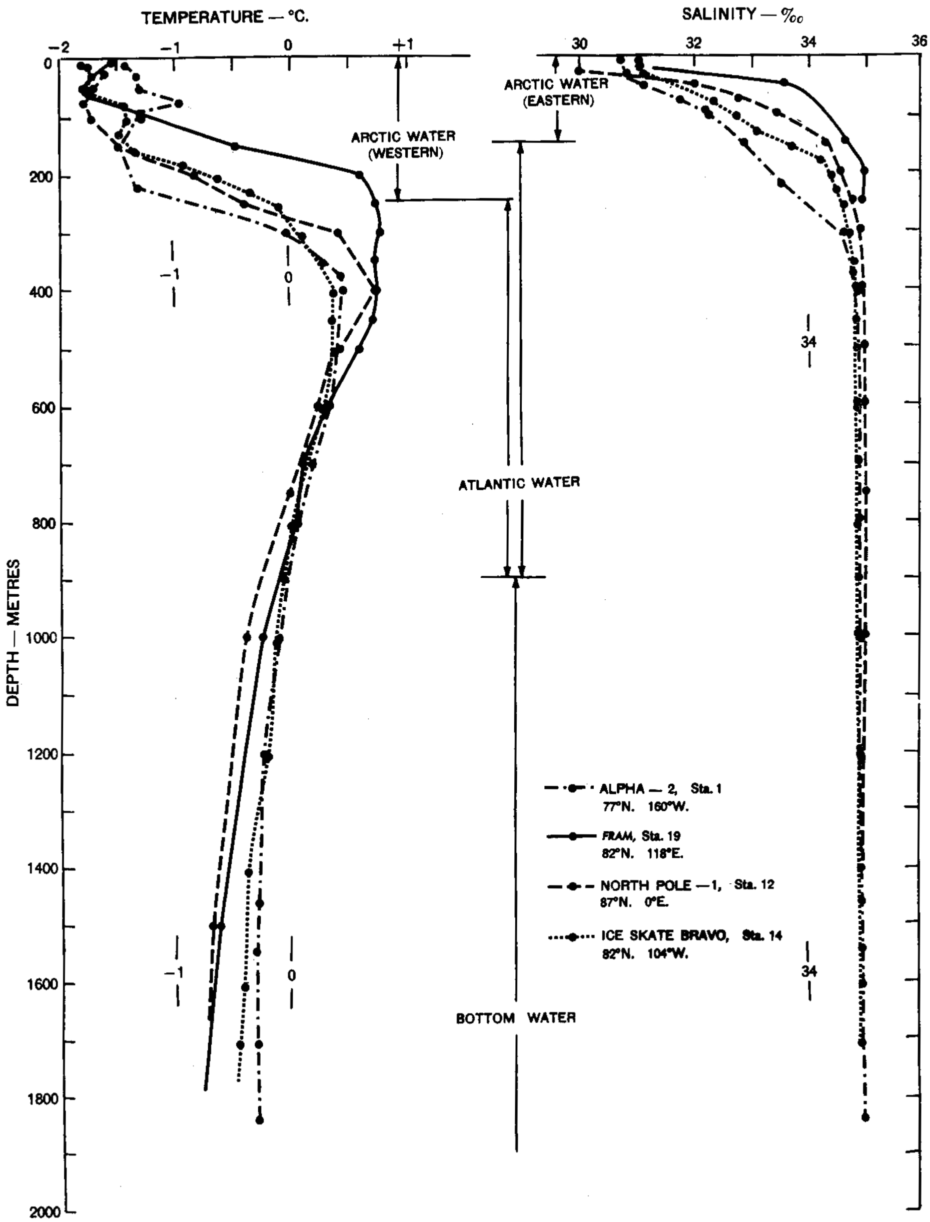


Fig. 2. Vertical distribution of temperature and salinity for four stations.

(2) The layer immediately below the Arctic Water, from about 150–250 metres down to 900 metres, has temperatures above 0°C. and quite uniform salinity (34.5–35.0 ‰). This water is undoubtedly of Atlantic origin (Nansen 1902, Timofeyev 1957a, 1958).

(3) Below this intermediate water layer lies bottom water with temperatures below 0°C. and with extremely uniform salinities between 34.93 and 34.99 ‰. This water is also of Atlantic origin, but apparently is formed only during winter and only in limited geographic areas in the Norwegian Sea (Sverdrup 1956, Timofeyev 1957b, Metcalf 1960).

In Fig. 2, two stations are from the eastern basin (*Fram,* Sta. 19 and North Pole-1, Sta. 12), and two from the western basin (Alpha-2, Sta. 1 and Ice Skate Bravo, Sta. 14). Even though the vertical distribution of temperature and salinity is generally similar throughout the Polar Basin, there is one notable difference between the two basins in the vertical temperature structure of the surface layer. The water in the western basin has a subsurface temperature maximum at 75–100 metres depth. This maximum may be strongly developed, with temperatures 0.5° to 1.0°C. higher than the water immediately above or below it (Alpha-2, Sta. 1), whereas in other areas it may be only barely discernible (Ice Skate Bravo, Sta. 14). This shallow temperature maximum was found in all seasons and years represented in the data from the western basin; there is, however, no evidence of its presence in the eastern basin, so apparently the degree to which it is developed depends on the locality.

In order to show that this shallow temperature maximum is indeed the effect of Bering Sea water, we must first discuss the circulation of Arctic Water in the western basin. The extensive current and ice-drift measurements made from the Russian drifting station North Pole-2 show that for the major part of the drift the ice and the water at 10 metres moved in a similar direction and with similar velocities. Hence the movements of the upper layers of water may be inferred from the movement of the numerous Russian and American drifting ice stations and of vessels such as the *Fram* (Fig. 3).

The drift of the ice island T-3, which has a deep draft (40 to 50 m. in contrast to 2 to 3 m. for ice floe stations), might be more nearly indicative of the circulation of a thicker layer of water because it is likely to be less influenced by direct wind stress. However, the drift of T-3 has been similar to that of other drifting stations in the same area, from which it may be inferred that the surface layers move as a unit. This is confirmed by the Russian current measurements previously cited.

The observed uniformity of temperature and salinity in time has allowed us to calculate the dynamic topography of the western basin (Fig. 4). The calculations were based on the 1200-decibar surface taken as the reference surface (assumed level of no motion), and were not carried to the sea surface but to the pressure surface of 25 decibars to avoid any seasonal influence affecting the very surface layers. The resulting topography agrees very well with the topography estimated by Worthington (1959) and with two topographies calculated by Russian workers from much more nearly synoptic data (Gudkovich 1959).

The circulation implied by the dynamic topography is in excellent agreement with the observed drift of ice, both as to direction and velocity. The upper layers circulate anti-cyclonically around the Beaufort Sea, with velocities of 1–5 cm./sec. around most of the gyral and with rather greater velocities (10 cm./sec.) immediately north of Alaska. Farther to the west (north of Siberia) north-flowing water does not necessarily enter the gyral but flows past the North Pole towards the east Greenland current.

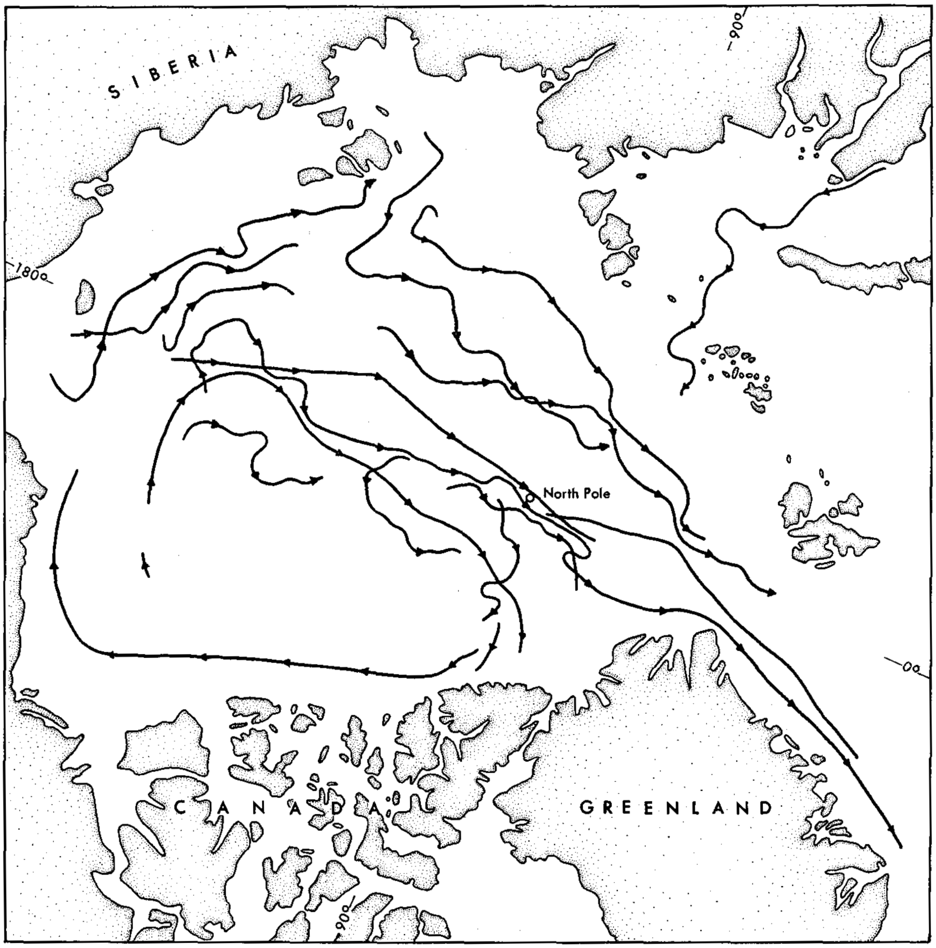


Fig. 3. Observed drift of floe stations and vessels in the Arctic Ocean. (*Source:* U. S. Navy Hydrographic Office. Publ. No. 705, 1958; Table 1.)

To examine the distribution and persistence of the shallow temperature maximum some of the data have been grouped, rather arbitrarily, as groups lettered A–K (Fig. 5). These groupings run clockwise in the Beaufort Sea, in keeping with the circulation of the Arctic Water (Fig. 1). An attempt has been made to include in each group data from more than one year. Fig. 5 discloses the following:

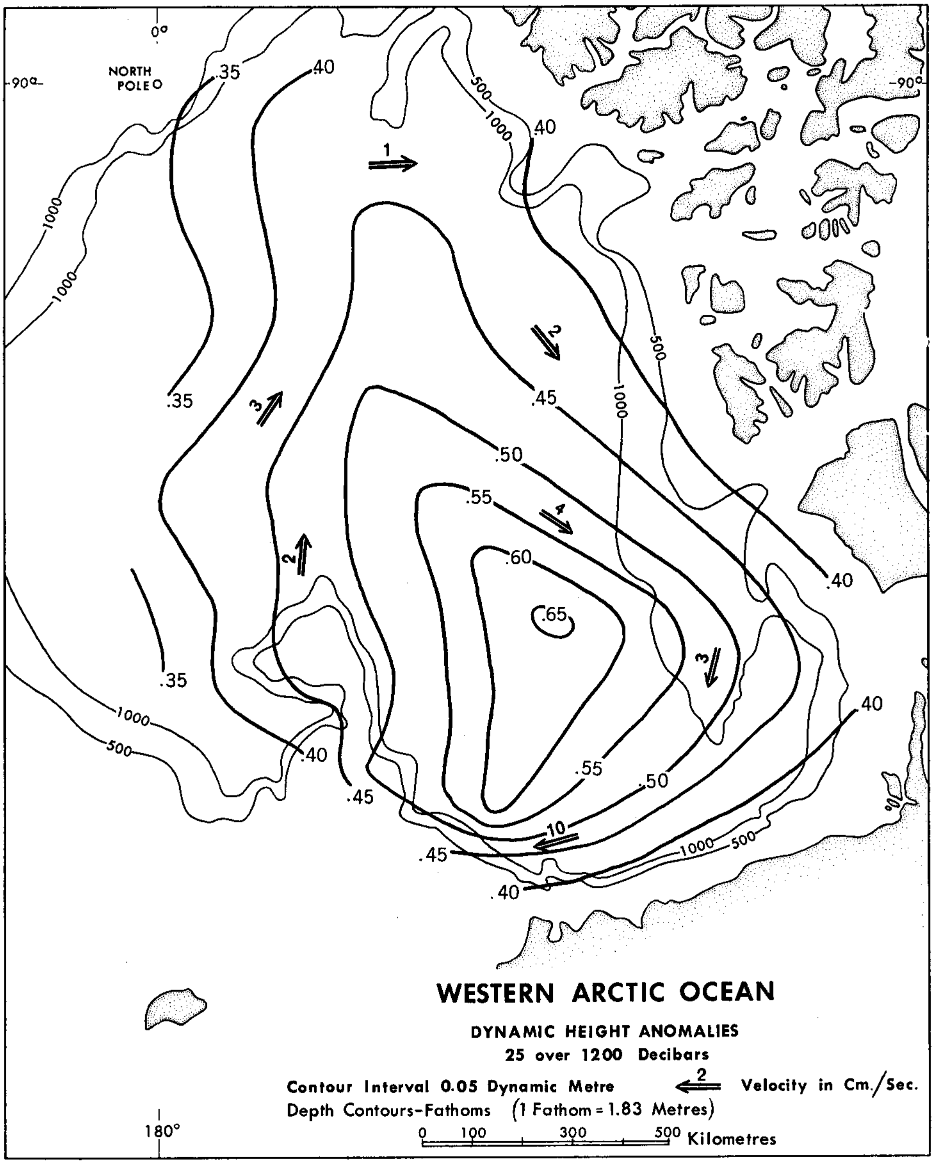


Fig. 4. Dynamic topography of the western Arctic Ocean.

(1) The shallow temperature maximum must be considered a persistent phenomenon because the curves are quite similar within each group even though the stations were occupied in different years and seasons.

(2) The shallow temperature maximum is best developed due north of the Chukchi Sea, and the difference between the maximum and ambient temperature decreases as the water travels around the Beaufort Sea gyral. The temperature maximum may be observed near the North Pole, as evidenced by the atypical station in G (the station in Fig. 1 located closest to the North Pole). It is present to some extent over the interior of the Beaufort Sea, and it occasionally may occur in an isolated locality north of Alaska (the atypical station in K).

(3) Below the maximum there is a temperature minimum at about 150 metres that is just as persistent as the maximum, with values from —1.4 to —1.5°C., even though the temperatures above and below are higher. Note that these temperatures are 0.2° to 0.3°C. above freezing.

It may be concluded that some continuing supply of relatively warm water is required to maintain the shallow temperature maximum. Such water either may acquire its characteristics, that is its temperature and salinity, locally or may acquire these at some distance and be advected into this part of the Arctic Ocean. The possibility of local origin is ruled out for the following reasons:

(a) The freezing of ice is the local process which would increase the density of the water to the point where it could sink to the required level. However, when this is due to freezing the temperature of the water leaving the surface is that of freezing for its salinity, whereas the water of the shallow temperature maximum is well above its freezing point.

(b) Heat might be supplied by radiation to raise the temperature above freezing, but the surface waters heated by radiation are of too low a density to sink even to 50 metres. This also rules out the possibility of the temperature maximum being formed and maintained by residual summer heat introduced in the relatively low-salinity, ice-free peripheral areas in the Arctic, e.g., north of Alaska, as suggested by Worthington (1953, 1959).

Thus, the shallow temperature maximum observed in the western Arctic Basin appears to be maintained by advection from some external source. The phenomenon is best developed in the area due north of the Chukchi Sea, and it is highly probable that the large amounts of water flowing north through Bering Strait into the Chukchi Sea contribute to some extent to the formation and maintenance of the temperature maximum. So long as it is present, then, the shallow temperature maximum would indicate the penetration of Bering Sea water into the Arctic Ocean. This view is substantiated by water-mass analysis.

Fig. 6 presents T-S diagrams in which a single observation of temperature and salinity is plotted as a point, and points of equal density show as curved lines labelled σ_t. The middle part of the diagram shows data from stations comprising groups A–D, the groups located due north

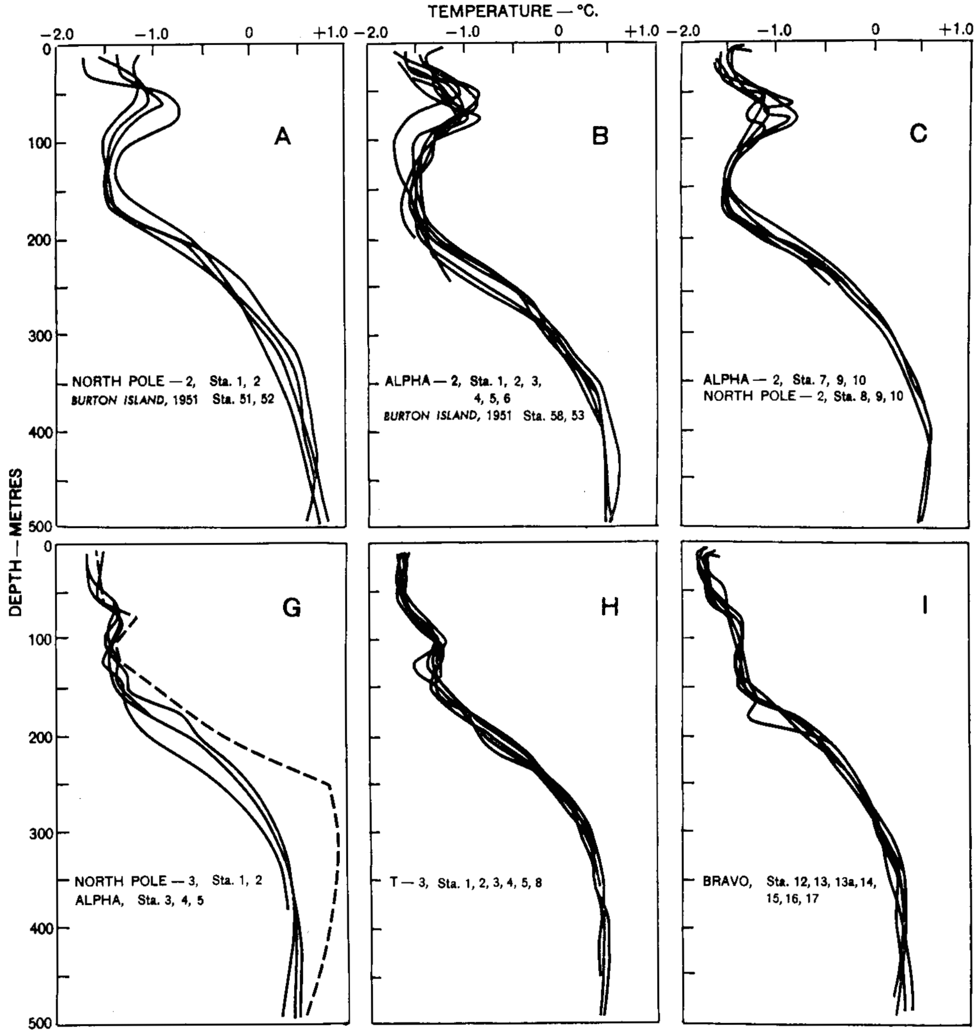
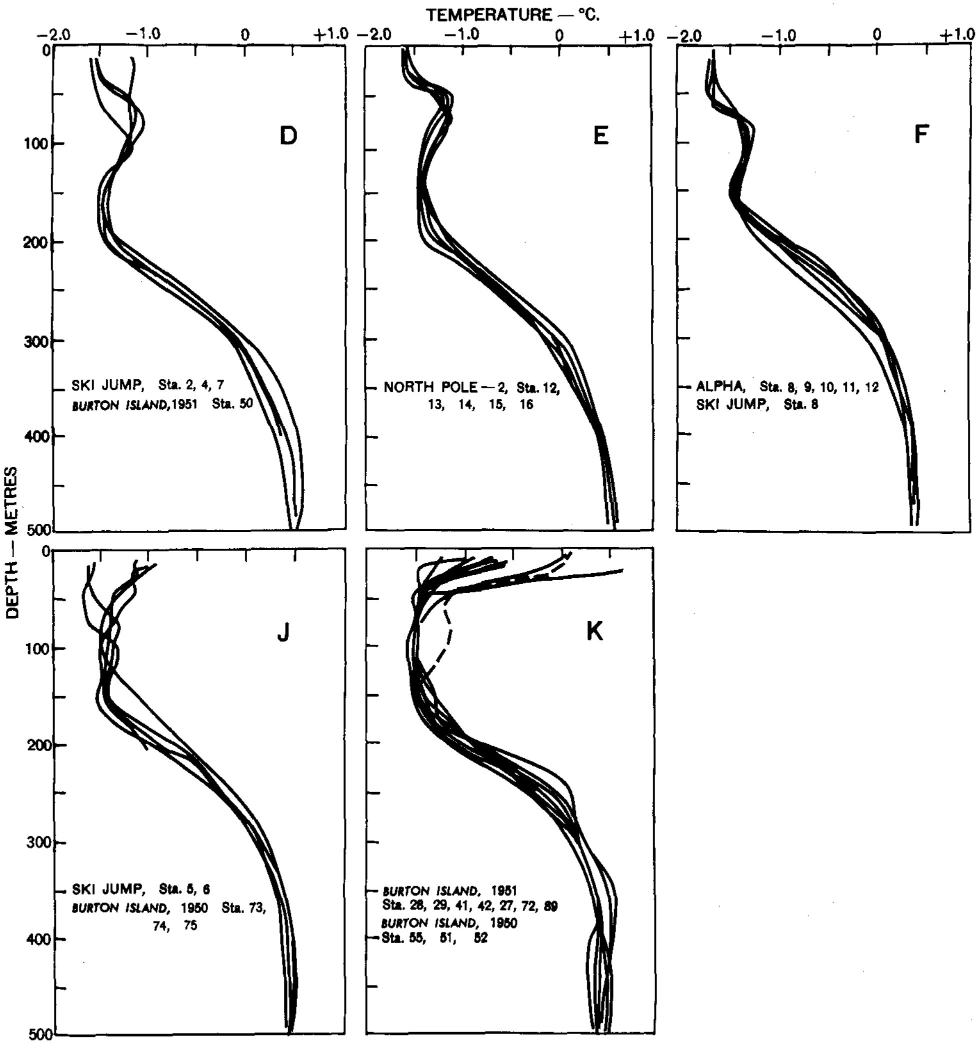


Fig. 5. Vertical distribution of temperature at selected groups of stations.

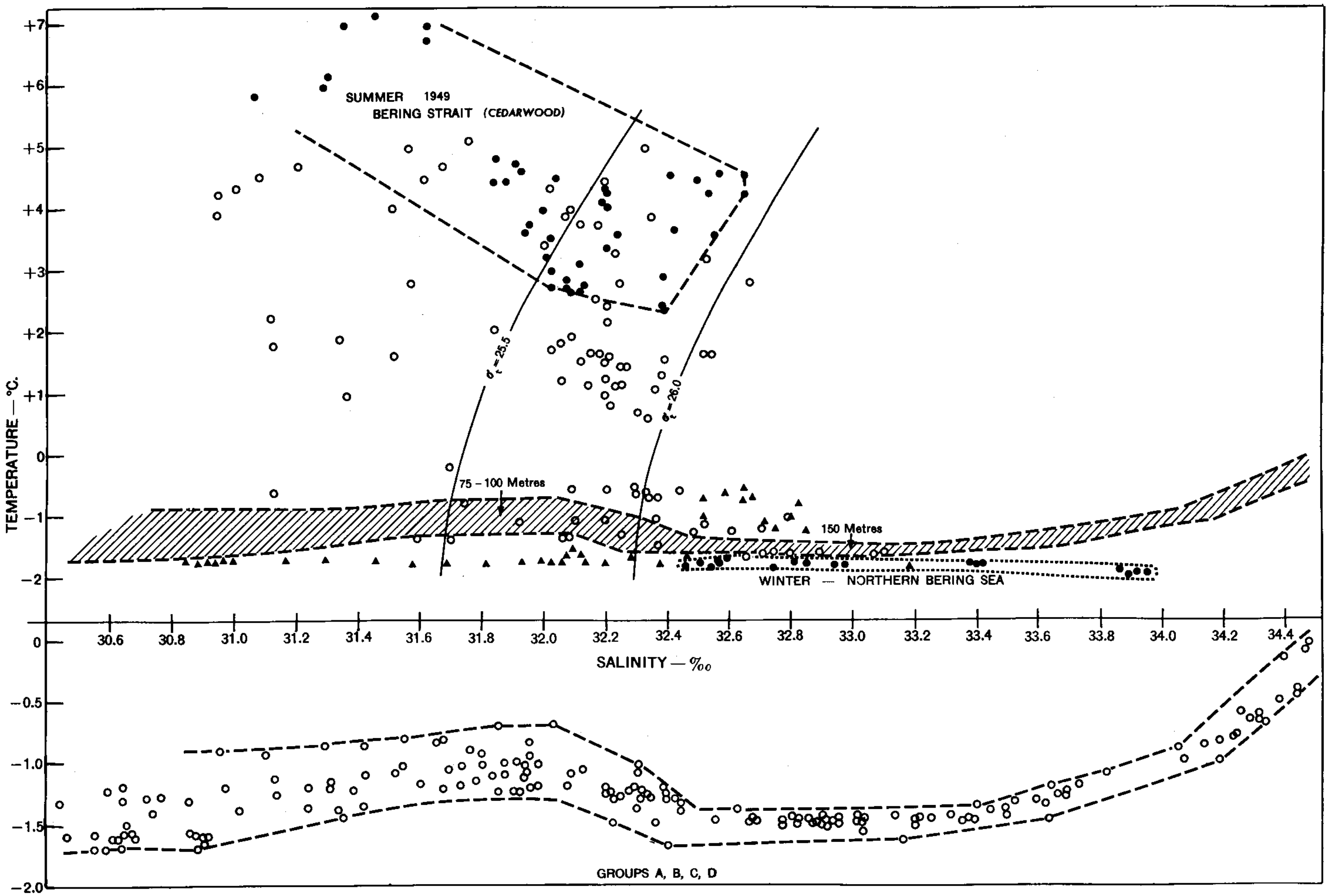
of the Chukchi Sea where the shallow temperature maximum was best developed. Note that the shallow temperature maximum appears as a "hump" in the salinity range 31.6–32.4 $^0/_{00}$. An envelope drawn around these values has been transferred to the upper and lower parts of the figure so that the water characteristics may be directly compared with those of the water flowing through Bering Strait and of the shelf water that occupies large parts of the Chukchi and East Siberian seas.

A number of stations are available from the western part of the Chukchi Sea from the *Maud* in 1922 (Sverdrup 1927). Individual observations are shown by triangles in Fig. 6. These data represent eastern Siberian shelf water. Even though taken in summer, the surface layers were cool because in 1922 the ice pack lay well to the south. Below the



(*See* Fig. 1 for locations of station groupings, Table 1 for sources of data.)

surface, in the level from 10 metres down to 30–60 metres, there was a fairly thick layer of intermediate shelf water with a salinity (and hence density) range the same as that of the water of the shallow temperature maximum. The temperature of this water was close to freezing, —1.5° to —1.6°C. According to Sverdrup, this intermediate water was encountered at all shelf stations and probably represented the water that during the previous winter had covered the greater part of the East Siberian and Chukchi seas. The bottom shelf water was of a different character, slightly saltier and warmer. It should be noted here that secular differences are not important in the use of *Maud* data, because we are discussing rather broad ranges of property; it is only necessary to assume that the processes involved have not changed significantly in the last 40 years.



TEMPERATURE — °C.
SALINITY — ‰
SUMMER 1949
BERING STRAIT (CEDARWOOD)
$\sigma_t = 25.5$
$\sigma_t = 26.0$
75 – 100 Metres
150 Metres
WINTER — NORTHERN BERING SEA
GROUPS A, B, C, D

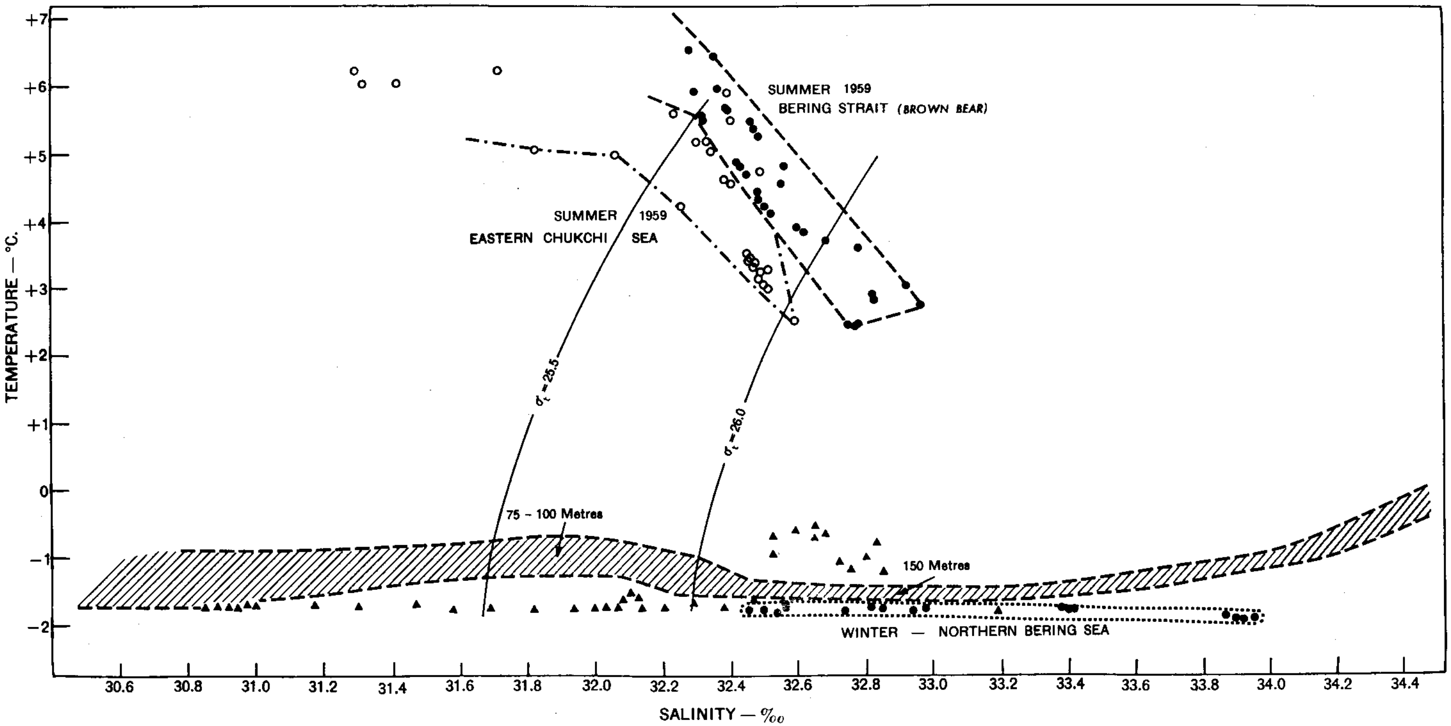


Fig. 6. Temperature-salinity observations from Bering Strait and eastern Chukchi Sea compared with observations from the Siberian shelf and Arctic Ocean. Upper — *Cedarwood* 1949; middle — observations from Arctic Ocean groups A — D from which the envelope in the upper and lower sections is derived; lower — *Brown Bear* 1959. Solid circles — observations from Bering Strait; open circles — observations from eastern Chukchi Sea; triangles — observations from *Maud* 1922. (*See* Table 1 for sources.)

What are the characteristics of the Bering Sea water that flows north through Bering Strait? Unfortunately, no concurrent temperature and salinity data in winter are available from Bering Strait proper, but during three icebreaker cruises in 1951 and 1955 stations were occupied just south of Bering Strait in the northern Bering Sea (U. S. Navy Hydrographic Office 1954, 1958). The data, plotted on the T-S diagrams, show that the water flowing north toward Bering Strait in winter may have a broad range of salinity (32.4–34.0 $^0/_{00}$) but is saltier than the water of the shallow temperature maximum. Its salinity range corresponds to that of the temperature minimum at 150 metres, and its temperature is that of freezing for its salinity.

The Bering Sea water flowing into the Chukchi Sea in summer is much warmer and also contains less salt than the winter water. Numerous data are available from a *Brown Bear* cruise in August 1959 (Fleming 1959) and a cruise of the *Cedarwood* in August 1949 (U. S. Navy Electronics Laboratory 1954). All observations made in Bering Strait during these two cruises are plotted in Fig. 6 and enclosed with a dashed line. At each station the observations were plotted from the bottom up, so that the coldest and most saline values are all included but warmer and fresher values nearer the surface may have been omitted.

The water flowing through Bering Strait in summer is considerably warmer and is somewhat less saline than in winter. It is significant that the resulting sigma-t (density) of most of this north-flowing water lies in the range 25.5–26.0, exactly the sigma-t range of the water of the shallow temperature maximum. Mixing of waters in the ocean may take place along sigma-t surfaces or across sigma-t surfaces, but on the basis of energy considerations the mixing that takes place is preferentially along surfaces of equal sigma-t (Sverdrup *et al.* 1942).

This analysis indicates that the water of the shallow temperature maximum originates as summer Bering Sea water, which, after flowing through Bering Strait, mixes in the Chukchi Sea mainly along sigma-t surfaces with the large amounts of intermediate shelf water. As further confirmation, observations made by *Brown Bear* and *Cedarwood* in the eastern Chukchi Sea considerably north of Bering Strait are also plotted in Fig. 6. As the water flowing north through Bering Strait tends to hold to the Alaskan coast, these observations lie in the path of this flow and it is apparent from the T-S diagrams that most of the mixing is taking place in the range of sigma-t 25.5–26.0.

Possibly the water of the temperature minimum at 150 metres may, in part, originate and be maintained by mixing of appropriate amounts of Bering Sea water with shelf water. In this case bottom shelf water would mix with winter Bering Sea water, and the resulting mixture would be denser and lie below the water of the shallow temperature maximum in the Arctic Ocean. The heat to maintain the water of the temperature minimum at 0.2° to 0.3°C. above freezing would be supplied by the bottom shelf water.

It is concluded that Bering Sea water continually flows north into the Chukchi Sea where it mixes with Siberian shelf water and then joins the Arctic Ocean circulation in the area northwest of Point Barrow. Estimating from the characteristics of the summer water, the water of the shallow temperature maximum is comprised of about 10–20 per cent Bering Sea water and 80–90 per cent intermediate shelf water. The consequences of the intrusion of this modified Bering Sea water on the oceanography of the Arctic Ocean may be summarized thus:

(a) The intruding Bering Sea water in the Beaufort Sea gyral effectively separates deeper Atlantic water from Arctic Ocean surface water, and also limits the depth to which local surface processes are effective in altering the properties of the water column. This separation severely limits any influence of Atlantic water on the surface water in this area, and it likewise limits the depth of local vertical convection associated with the freezing of ice. The deepest convection found as evidenced by homogeneous properties is to depths of less than 70 metres.

(b) The shallow temperature maximum is an excellent tracer of the Bering Sea water, and makes it possible to estimate the vertical eddy coefficients associated with its decay as the water travels around the Beaufort Sea gyral. For the grouped temperature profiles, the value of the temperature maximum may be considered a result of the balance of horizontal advection by vertical diffusion:

$$U\frac{\partial \theta}{\partial x} = \frac{\partial}{\partial z}\left(K_z\frac{\partial \theta}{\partial z}\right)$$

where
U = average velocity
θ = temperature
x = distance along the line of flow
z = distance in the vertical
K_z = vertical eddy coefficient

Using data from drifting-stations where the ice tends to follow the same water mass, the balance may be simply expressed by :

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(K_z\frac{\partial \theta}{\partial z}\right)$$

where t = time. Solutions of these equations for K_z lead to values of between 0.2 and 1.1 $cm.^2\ sec.^{-1}$.

(c) The inflowing water from Bering Sea will have very little influence on ice conditions in the Arctic Basin. The winter water is at the freezing point, the same as the upper and intermediate water in the Chukchi Sea. The summer water loses its heat very rapidly through mixing with the cold intermediate shelf water, and the density of the mixture is such that it enters the Arctic Ocean at subsurface levels. Thus the influence of Bering Sea water on the ice cover will be limited to the upward flux of heat from the water of the shallow temperature maximum to the very surface layers (upper 50 m.) where it could affect the ice. The maximum upward heat flux is estimated by using the observed vertical temperature distribution and the calculated eddy coefficient. Solution of

$$\frac{1}{A}\frac{dQ}{dt} = \rho\, C_p K_z \frac{d\theta}{dz}$$

where A = area, Q = heat, ρ = density and C_p = specific heat gives a maximum value for the heat flux upward across a 1 cm.^{2} area at 50 metres depth of 4 x 10^{-4} cals. cm.$^{-2}$ sec.$^{-1}$ or about 35 cals. per cm.^{2} per day. This calculated upward flux of heat is considerably less than the estimated heat loss from the surface in these high latitudes, which is about 330 cals. per cm.^{2} per day gross or 108 cals. per cm.^{2} per day net (Sverdrup *et al.* 1942, p. 99).

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