SURFACE WATER IN THE EURASIAN BASIN OF THE ARCTIC OCEAN*

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THE Arctic Ocean can be separated, on the basis of temperature and salinity characteristics of the water, into a three-layered system. This layered nature of the water in the Arctic Ocean was first observed by Nansen (1902) who, during the drift of the *Fram*, noted that the top layer of the sea, between 0 and 200 metres consisted of water of low salinity and low temperature; this layer was termed by him "genuine polar water". From 250 m. to the bottom the sea was filled with water of very high salinity and relatively high temperature, which evidently originated in the Atlantic Ocean and was slightly modified on the way to the Arctic Ocean. The temperature of the deeper water was above 0°C. down to about 900 m. and less than 0°C. below 900 m. but not so cold as the upper polar water.

Studies have been conducted since the expedition on the *Fram* with ever increasing intensity. These studies have not altered the basic concept of the physical oceanography of the Arctic Ocean as elucidated by Nansen, but they have added much detail. The considerable amounts of station data that have been accumulated over many years, and during all seasons of the year, show a remarkable regularity of the distribution of temperature and salinity at depth throughout the year, as well as repeating seasonal regularities in the surface waters. [It should be noted that in waters as cold as those in the Arctic Ocean the vertical distribution of density so closely parallels that of salinity that the latter may be used as an index of the mass distribution.] This close approach to a steady state in the observed distribution of these properties is apparently the result of continuing processes within the basin. Because of this the layered waters of the Arctic Ocean can be divided into water masses with defined characteristics of temperature and salinity, and analyses can be made of their origin and distribution without recourse to synoptic data.

The Lomonosov Ridge divides the Arctic Ocean into two major deep-water basins, the Canadian and the Eurasian, as shown in Fig. 1a (LaFond

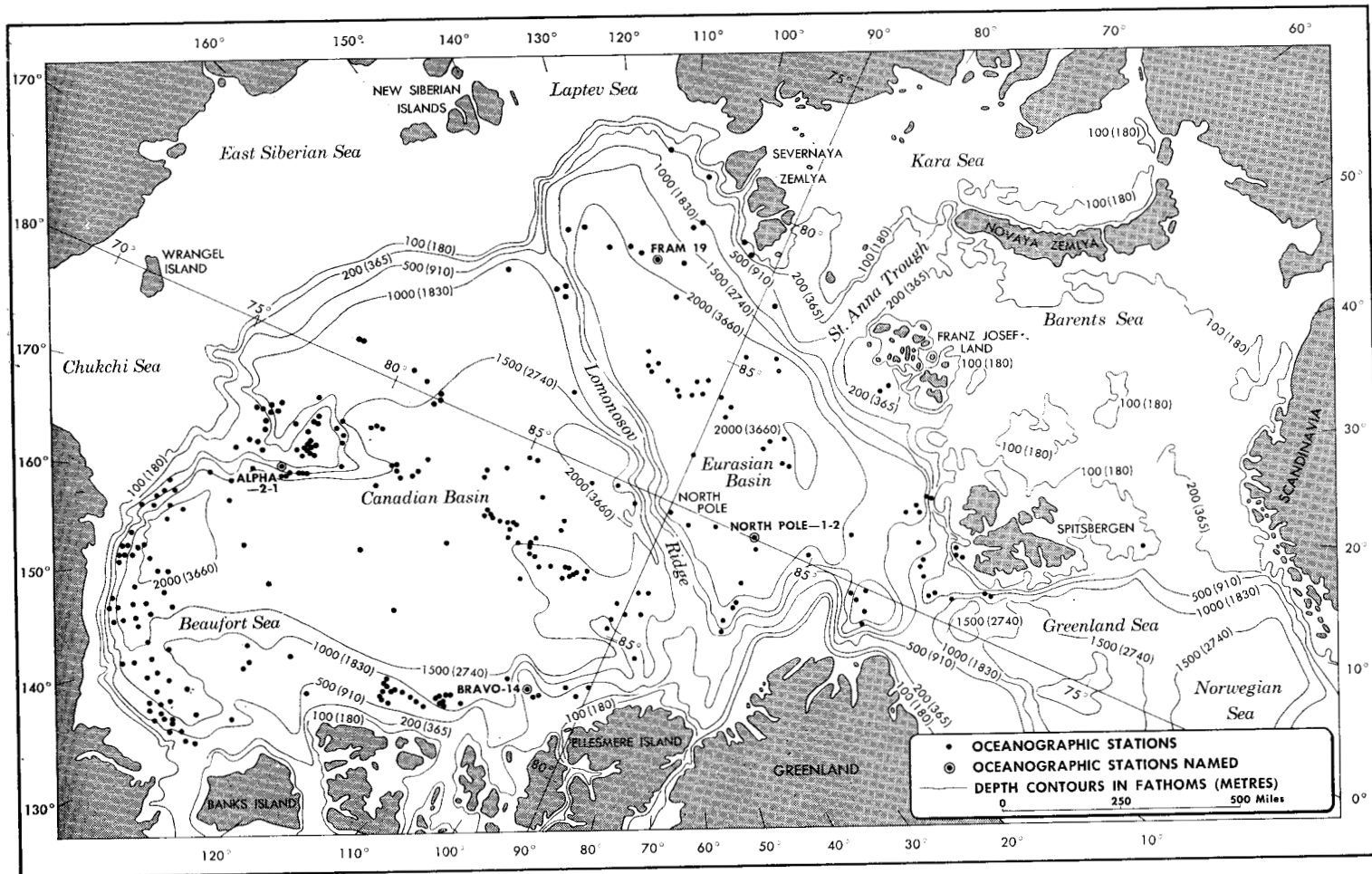† Dept. of Oceanography, University of Washington.

Fig. 1a. The Arctic Ocean.

1960). In a previous paper (Coachman and Barnes 1961), we have shown that the surface water of the Canadian Basin contains a slight temperature maximum in the lever 50-100 m. and a temperature minimum below it at 150 m. The maximum was shown to be a result of the influx of Bering Sea water in summer through Bering Strait, whereas the minimum is related to the winter inflow. The temperature of this contribution of water from the Bering Sea dominates the temperature structure of the surface water over much of the Canadian Basin, but there was no evidence of any influence of this water on the surface water of the Eurasian Basin.

In the Eurasian Basin, in spite of the absence of a contribution of Bering Sea water to that overlying the Atlantic layer, the surface water occupies a layer nearly as thick as that in the Canadian Basin, that is from 0 to 200 m. It is generally considered that this relatively thick surface layer is basically a mixture of continental runoff water that flows into the central basins in modified form from the peripheral seas with water of high salinity (cf. Nansen 1902, Shirshov 1944, Sverdrup 1950, 1956, Treshnikov 1959a, Timofeyev 1960). The annual volume of continental runoff (Antonov 1958) is one-tenth that of the inflow through Bering Strait and one-fiftieth that of the outflow via the East Greenland current, and is added primarily to the Barents, Kara, and Laptev seas, which border the Eurasian Basin. The question is where and with which water does the runoff water mix to exercise an important influence on the surface water in the Eurasian Basin.

An additional problem is a peculiarity that has been observed in the vertical distribution of temperature and salinity in the surface water. Nansen (1902), working with data from the *Fram* drift, noted that the halocline separating the surface water of relatively low salinity from the more saline Atlantic layer began at about 30 m. and was very strongly developed at 50-60 m., whereas the temperature, which was low near the surface (close to the freezing point), continued to decrease and even exhibited a minimum at 60 m. He suggested that this under-surface layer was advected into the region of the *Fram* drift from another part of the basin where the surface of the sea would be saltier and, being at the freezing point, would therefore have a slightly lower temperature.

Shirshov (1944), working with the data from station North Pole-1, observed this same distribution of temperature and salinity, and he divided the surface water into two water masses: the layer from near-surface down to 25-50 m. with a uniform salinity, and the layer from 50 m. to the Atlantic layer with a lower temperature and a sharply increasing salinity. The latter he named "intermediate water" and considered it to be formed by the mixing of surface arctic water with Atlantic water.

Timofeyev (1951, 1960), working with data from the *Sedov* drift, made a different division of the surface water: the layer from the surface to 50-100 m. with a uniform temperature and a salinity increasing with depth, the "surface arctic water mass", and the layer from 50-100 m. to 200-250 m. with a temperature rapidly rising to 0°C. and a gradually increasing salinity, the "intermediate water mass". The latter he reported to be formed as a result
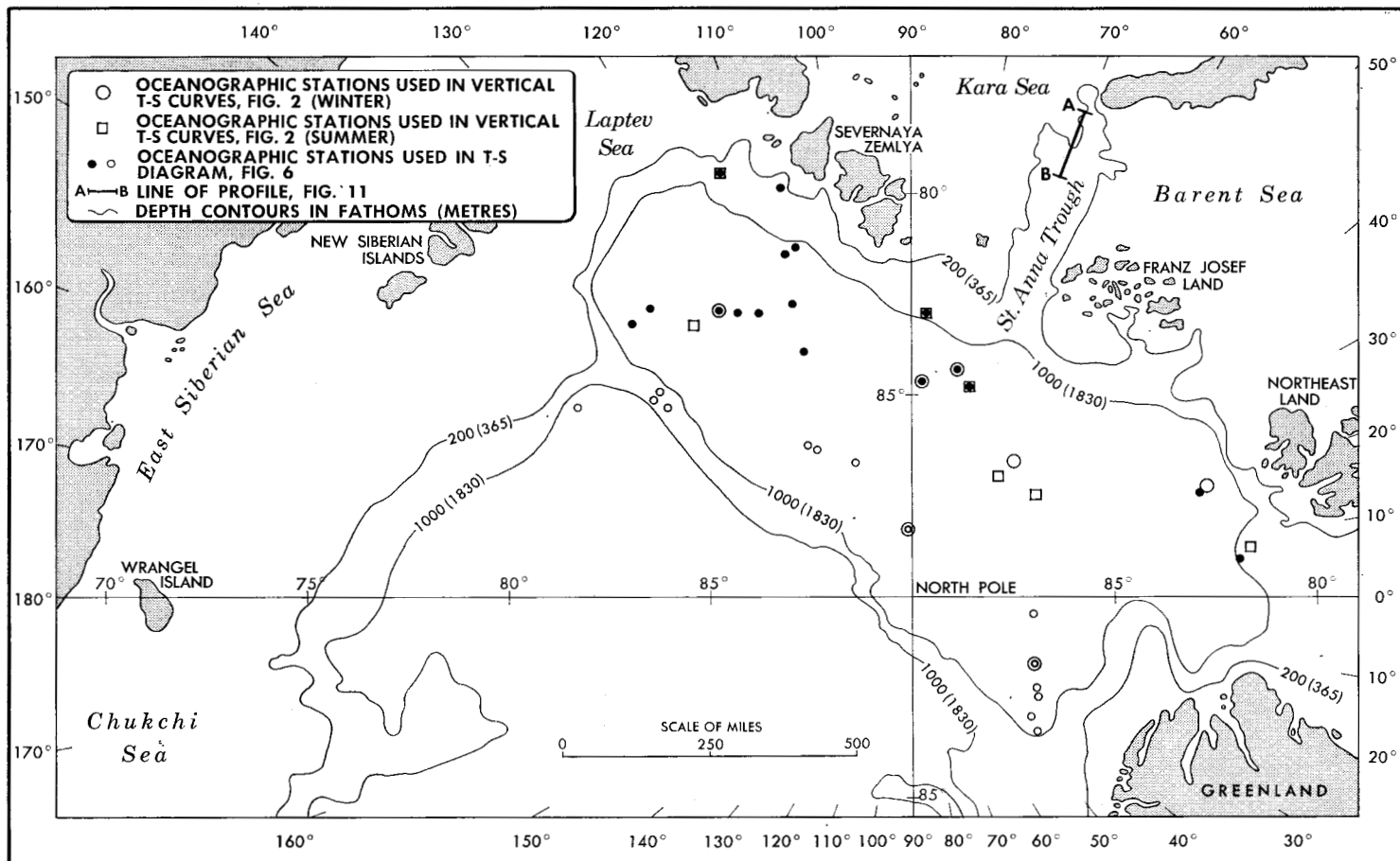
Fig. 1b. The Eurasian Basin, showing locations of the stations used in Figs. 2 and 6, and the profiles of Fig. 11.

of a gradual mixing of surface arctic water with underlying Atlantic water. He considered the slight temperature minimum noted by Shirshov to be residual, the near-surface layers being warmed slightly in summer.

Treshnikov (1959b) pointed out that the 50- to 100-m. layer, with the same (or slightly lower) temperature as the surface layer but sharply increasing salinity, could not result from mixing of the surface water with Atlantic water because if that happened the temperature would also increase. He related this subsurface water to the surface water of high salinity ($>33.5\%_0$), which is formed by the freezing process and mixed by vertical convection in winter and which occupies a strip from Greenland eastward along the continental slope to the Laptev Sea. He supposed that this water, with a density greater than that of the surface water throughout the rest of the basin, gradually spreads to the north, sinking to 100 m. and carrying its low temperature with it.

With the help of a grant from the Arctic Institute of North America we have undertaken to re-evaluate, analyze, and interpret all available data from the Arctic Ocean. This paper reports the result of our interpretation of data from the surface-water layer of the Eurasian Basin.

Temperature and salinity

More than 400 oceanographic stations have been occupied in the Eurasian Basin (Gordienko and Laktionov 1960), but only 74 of these are available for examination (Table 1). Fig. 2 presents 14 of these stations, distributed

Table 1. Oceanographic stations used in this analysis.

Vessel or station	Number of stations	References
Stations from the deep-water basin:		
Fram (1894)	9	Nansen 1902
Nautilus (1931)	7	Sverdrup 1933
Sibiryakov (1932)	2	Vize 1933
Sedov (1937-40)	9	Timofeyev 1960
North Pole-1 (1938)	6	Shirshov 1944
North Pole-1 (1938)	3	Timofeyev 1960
North Pole-3 (1954)	2	Treshnikov and Tolstikov 1956
North Pole-3 (1955)	3	Musin 1960
North Pole-4 (1957)	1	Timofeyev 1960
North Pole-5 (1955)	5	Timofeyev 1960
North Pole-6 (1958)	4	Tomifeyev 1960
North Pole-7 (1958)	1	Timofeyev 1960
Expedition on Drifting Ice 1951	1	Timofeyev 1960
High-Latitude Air Expedition 1948	1	Timofeyev 1960
High-Latitude Air Expedition 1948	1	Musin 1960
High-Latitude Air Expedition 1949	1	Timofeyev 1960
High-Latitude Air Expedition 1950	1	Timofeyev 1960
High-Latitude Air Expedition 1955	10	Timofeyev 1960
Litke (1948)	1	Gur'yanov and Musin 1960
Atka (1954)	2	U.S.N.H.O. 1954
Litke (1955)	1	Timofeyev 1960
Unknown (1935, 1948, 1949)	3	Musin 1960

WINTER (NOV.-MAY)

○	FRAM STA. 14	(80°N. 132°E.)
●	FRAM STA. 22	(85°N. 87°E.)
□	H.I.A.E. 55 STA. 56	(84°N. 78°E.)
■	SEDOV STA. 67	(86°N. 54°E.)
△	UNKNOWN (1949)	(88°N. 90°E.)
▲	H.I.A.E. 55 STA. 32	(82°N. 21°E.)
×	NP-3 STA. 19	(87°N. 28°W.)

SUMMER (JULY-SEPT.)

○	LITKE 1948 STA. 146	(78°N. 114°E.)
●	FRAM STA. 18	(81°N. 128°E.)
□	UNKNOWN (1935)	(83°N. 87°E.)
■	FRAM STA. 23	(85°N. 75°E.)
△	SEDOV STA. 69	(86°N. 56°E.)
▲	SEDOV STA. 72	(86°N. 41°E.)
×	LITKE 1955 STA. 30	(82°N. 9°E.)

DEPTH IN METRES
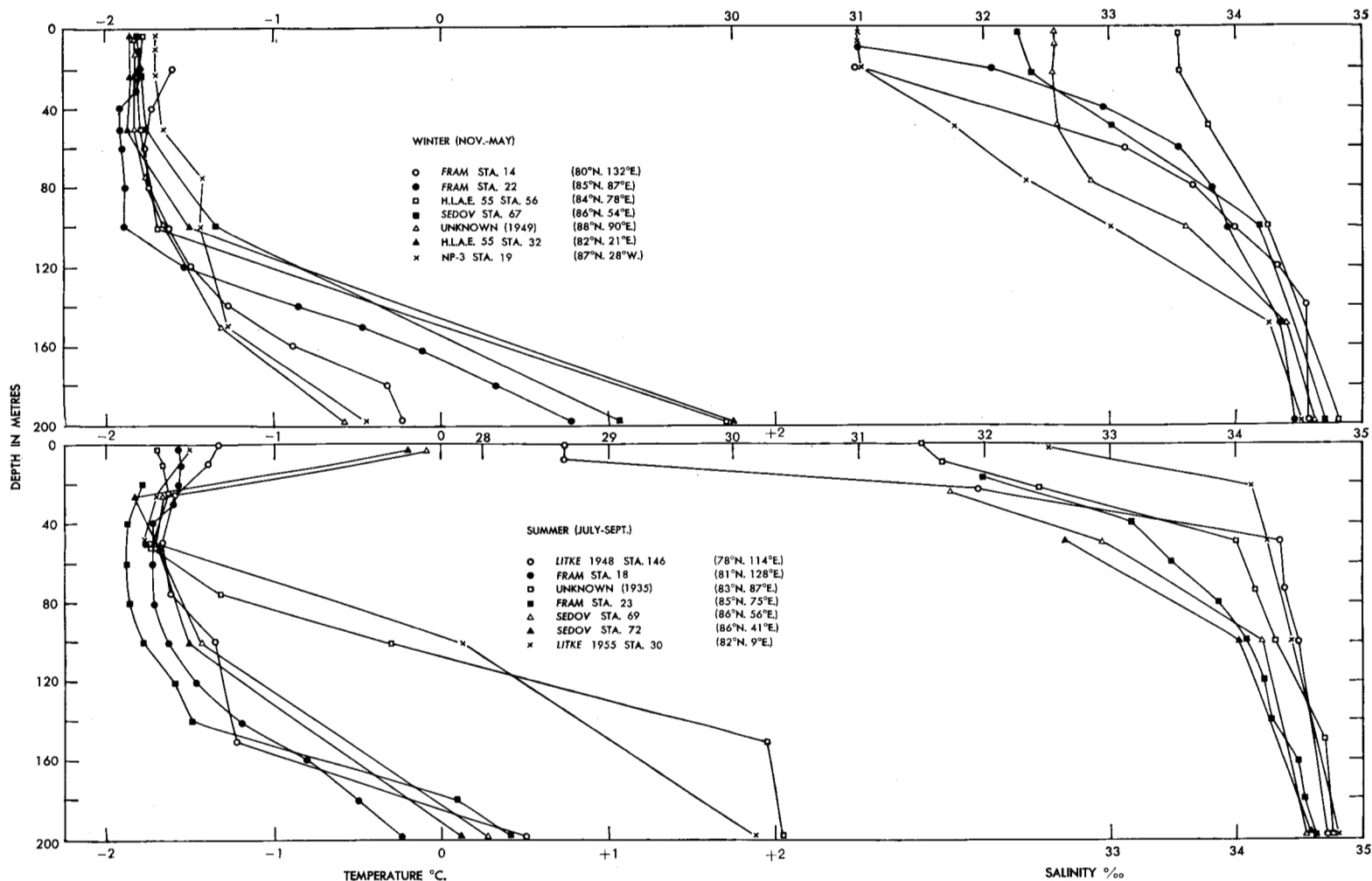
TEMPERATURE °C.

SALINITY ‰

Fig. 2. Vertical distribution of temperature and salinity for 14 stations.

Table 1. *(continued)*.

Sea	Vessel	Number of stations	References
**Stations from the northern fringes of the peripheral arctic seas:**		
Barents	*Sedov* (1929)	14	Vize and Laktionov 1931
	Quest (1931)	79	Mosby 1938
	Taimir (1934)	9	Laktionov and Balakshin 1936
	Litke (1948)	5	Gur'yanov and Musin 1960
Kara	*Lomonosov* (1931)	22	Laktionov 1935
	Sedov (1934)	70	Laktionov and Balakshin 1936
	Sadko (1935)	2	Gur'yanov and Musin 1960
	Unknown (1935)	2	Musin 1960
	Litke (1948)	6	Gur'yanov and Musin 1960
Laptev	*Rusanov* (1932)	23	Arkticheskii Institut 1936
	Sibiryakov (1932)	17	Vize 1933
	Krasin (1933)	5	Maksimov 1936
East Siberian	*Maud* (1922-4)	93	Sverdrup 1929

over the length of the basin, to illustrate the vertical distribution of temperature and salinity (locations shown in Fig. 1b). These stations have been selected to give not only an adequate areal coverage of the basin but also a representation of the distribution in summer and winter.

The surface layer (0-25 m.) is cold and relatively dilute. The salinity has a relatively broad range from 28.5 to 33.5‰, but there is some tendency for the values to be lower in summer (28.5-32.5‰) and higher in winter (31.0-33.5‰). The whole layer is always nearly isohaline down to 25 m. The temperature is always less than 0°C. and mostly close to the freezing point of the water. In winter this is strictly true; the layer is isothermal and the temperature ranges only from −1.60 to −1.85°C. In summer the temperatures are apparently slightly above freezing, particularly at the 5-m. level, and have a much broader range (−1.7 to 0°C.).

Below 25 m. the water continues to be nearly isothermal down to approximately 50 m. and often to 100 m., whereas the salinity increases markedly throughout the same depths. The increase of salinity with depth may be as much as 0.08‰ per metre or as little as 0.008‰ per metre, but at 100 m. all salinities are over 33‰ and some greater than 34‰.

At 100 m. the range in temperature is about 2°C. and in salinity about 1.5‰, with an apparent tendency for the water at this level to be warmer (−1.8 to +0.1°C.) and more saline (34.0 to 34.5‰) in summer than in winter (−1.9 to −1.3°C., 33.0 to 34.4‰).

The outstanding thermal feature of the Arctic Ocean begins below 100 m. The temperature rises from the low values near the surface to those of the core of Atlantic water (+0.7° to >+2.0°C., depending on locality), which lies below 200 m. At the same time the salinity increase becomes smaller than 0.002‰ per metre as the salinity approaches that of the Atlantic layer (34.9-35.0‰).

Discussion

The general pattern of distribution of temperature and salinity as described by the various authors cited is apparent. The surface water itself may be divided into layers: the near-surface to 25-50-m. layer, cold and dilute and subject to considerable seasonal fluctuations; the layer between 25 to 50 m. and 100 m. with a more or less uniform temperature but a sharp increase in salinity; and the layer below 100 m. in which the temperature and salinity values are transitional between the values of the layers above and those of the Atlantic layer.

In the observed values and in the vertical distributions there are some variations that account for the discrepancies in description between authors who worked mainly with results from only one expedition: e.g., between Shirshov 1944 (North Pole-1) and Timofeyev 1951 *(Sedov)*. These variations could be due either to (i) the variations associated with geographic localities within the basin that arise because the inflowing water enters only at specific sites, or (ii) long-term trends or seasonal fluctuations affecting the water entering the basin, which would lead to differences in the temperature and salinity structure even though the oceanographic processes within the basin tend to produce a steady state distribution.

Seasonal fluctuation has been shown in the Atlantic water where a synoptic survey, the High-Latitude Air Expedition of 1955, disclosed a wave in the temperature of the core of Atlantic water with a range of 1°C. (Timofeyev 1958).

Table 2. Temperature (°C.) at 200 metres.

1939	1955	Difference (1955 – 1939)
—0.20 (*Sedov* No. 57)	+0.04 (NP-5* No. 18)	+0.24
—0.03 (*Sedov* No. 46)	+0.26 (NP-5 No. 16)	+0.29
+0.13 (*Sedov* No. 72)	+0.30 (H-LAE† 1955 No. 27)	+0.17

*NP = North Pole
†H-LAE = High-latitude air expedition

Variations in temperature and salinity might also arise from secular changes in climate, which would affect primarily the temperature of the relatively warm Atlantic inflow and the salinity of modified continental runoff. The physics of the freezing process would sharply limit the range of temperature fluctuations of the latter inflow. If some of the observed variations are secular, they cannot be distinguished from seasonal changes in the present data. For example, stations separated by less than 30 km. but by 16 years in time (Table 2) show temperature differences at the 200-m. level of 0.2 to 0.3°C., whereas the range of temperature at 200 m. from place to place in the basin may be ten times this amount. Furthermore, the vertical temperature increases at this depth are such that a vertical displacement of less than 10 m., through inaccurate positioning or internal waves, could lead to

temperature differences as great as 0.3°C. We conclude that long-term trends in temperature and salinity values of the surface water, unless they are of greater magnitude than they apparently have been in the last 60 years, will to a large extent be obscured by seasonal fluctuations and by variations with location in the basin.

We will now consider the various layers of the surface water in some detail.

Layer from surface to 25-50 m. The deep-water part of the Arctic Ocean is covered with a more or less continuous ice blanket, which, on the average, is 3-3.5 m. thick at the end of winter and 2-2.5 m. thick at the end of summer (Defant 1961), but which may vary locally from open water leads to rafted areas of greater than average thickness (LaFond 1960). The presence of the ice cover is responsible for maintaining the temperature of the near-surface water at or close to its freezing point summer and winter. On the other hand, the irregularity in the ice cover can cause variations in temperature and salinity values such that a surface observation is representative only of the sampled spot. In winter the freezing of leads can locally increase the salinity of water parcels and in some instances water can be supercooled and the formation of ice can follow a decrease in temperature by a small time interval (Dorsey 1940). These effects can produce instability in the water column by increasing the density of the water parcel above that of the ambient surface or immediately underlying water.

In summer meltwater collects on the surface of open leads or holes in the non-uniform ice cover so that quite low values of salinity can obtain in the hole or lead. The meltwater is fresh and at its freezing point, 0°C., and thus this water not only reduces the salt concentration of the ambient water but also adds heat. In the openings, radiant energy can be absorbed directly by the water, also causing a temperature rise. The resulting fresher and warmer water parcels are less dense than the ambient surface water and tend to remain at the surface. However, due to the irregularity of the ice cover, differential movement of ice and water can mix the surface water down to about 5 m., which would tend to produce homogeneity in temperature and salinity. The net result is that observations at 5-m. depth are probably more nearly indicative of general surface conditions than surface observations, but even at this level the temperatures show some scatter around the freezing point. The freezing-thawing processes give rise to more or less continuous small local changes in density, and these apparently are compensated for rather slowly within the water column.

Examination of the 5-m. observations in Fig. 2, considered to be most nearly representative of ambient surface conditions, shows the salinity at this level to have a broad range, about 4‰, while the temperature has a narrower range (<2°C.). The reason for the narrow range of temperature is the limit imposed by the freezing process (see below). The range observed in salinity is probably at least partly due to the horizontal gradient of salinity that has been observed in the surface layers of the Eurasian Basin. Treshnikov (1959b) presented a chart of the average salinity of the 0- to 25-m.
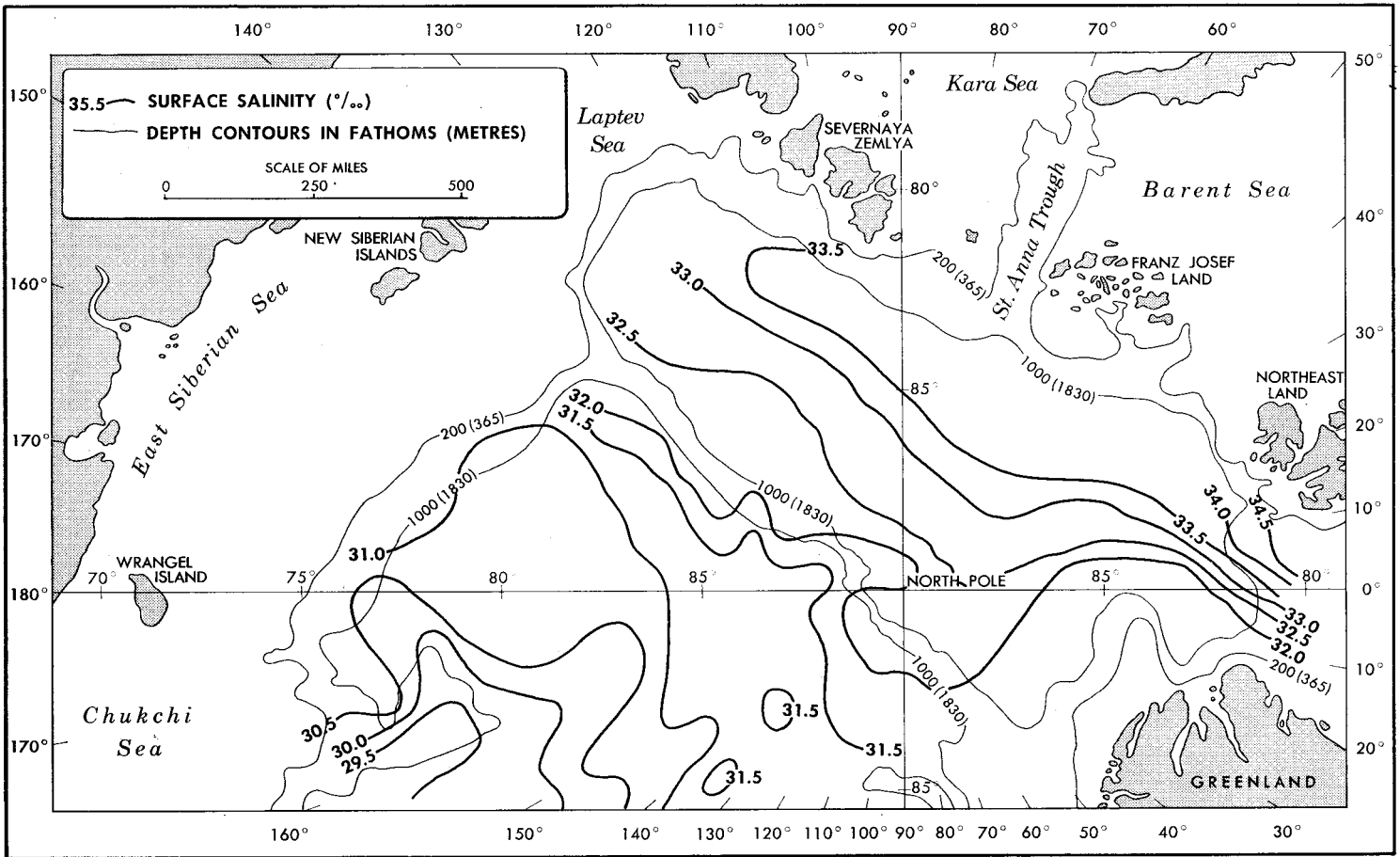
Fig. 3. Values of surface salinities in winter (after Treshnikov 1959b).

layer for the winter period (Fig. 3), based on numerous values from the High-Latitude Air Expeditions. The highest upper-layer salinities, greater than 33.5‰, occur in a strip approximately 500 km. wide bordering the continental shelf from Spitsbergen eastward to north of the Laptev Sea. Bands of water of lower salinity parallel this band on the northern side, and the strip from the New Siberian Islands toward Greenland has a salinity of 32-32.5‰. Observations from vessels or stations that drifted across the basin agree well with the picture presented by Treshnikov, e.g., Nansen (1902) found that the salinity increased on the average nearly 2‰ in 1000 km. going from east to west.

In nearly closed basins in higher latitudes there will be a surface outflow of fresh water with some saline water entrained with it and an inflow of more saline water at depth to maintain the salt balance (Sverdrup *et al.* 1942). In such basins there will be an increase in surface salinities from the head of the basin, close to the source of fresh water, toward the mouth due to progressive mixing between the surface layer and the underlying more saline water. In the Eurasian Basin the drift of vessels and drifting stations have indicated that, in general, the surface water and ice tend to flow on the shortest course from the peripheral seas to the basin exit between Spitsbergen and Greenland (Gordienko 1961), whereas the Atlantic water flows in nearly the opposite direction (Timofeyev 1957). Thus, the general conditions for a basin in high latitudes are fulfilled by the Eurasian Basin; the inflow of saline Atlantic water is largely confined to the strait between Spitsbergen and Greenland, and most of the fresh water input is spread along the Siberian coast, where it enters the Arctic Ocean via the peripheral seas. However, the variation in surface salinity does not coincide with the distribution of fresh water sources. More than one-half of the annual inflow of continental water is concentrated in the Kara and Laptev seas (Antonov 1957), but it is adjacent to these seas where the highest values of surface salinity are observed. This means that the north-flowing surface water in the northern parts of these seas has a relatively high salinity, which in turn implies either that the large quantities of fresh water are here more intensely mixed with even larger volumes of saline water than in the other peripheral seas, or that the surface waters are concentrated to a greater extent within these seas. However, the concentration of salt in the Arctic occurs through the freezing of ice, primarily in winter, and there is no evidence that significantly more ice forms in the Kara and Laptev seas than anywhere else in the Polar Basin; on the contrary, the Kara Sea has been observed to contain unusual areas of open water in winter (Kupetskiy 1959, Eskin 1960). We must conclude that the processes conditioning the surface water are not the same in all the bordering seas, and in the next section a model, which could account for the observed distribution of properties, is proposed.

Superimposed on the spatial variation in temperature and salinity are noticeable seasonal effects (cf. Fig. 2), but what is their magnitude and to what depth are they effective? Nansen (1902) recorded that the effect of
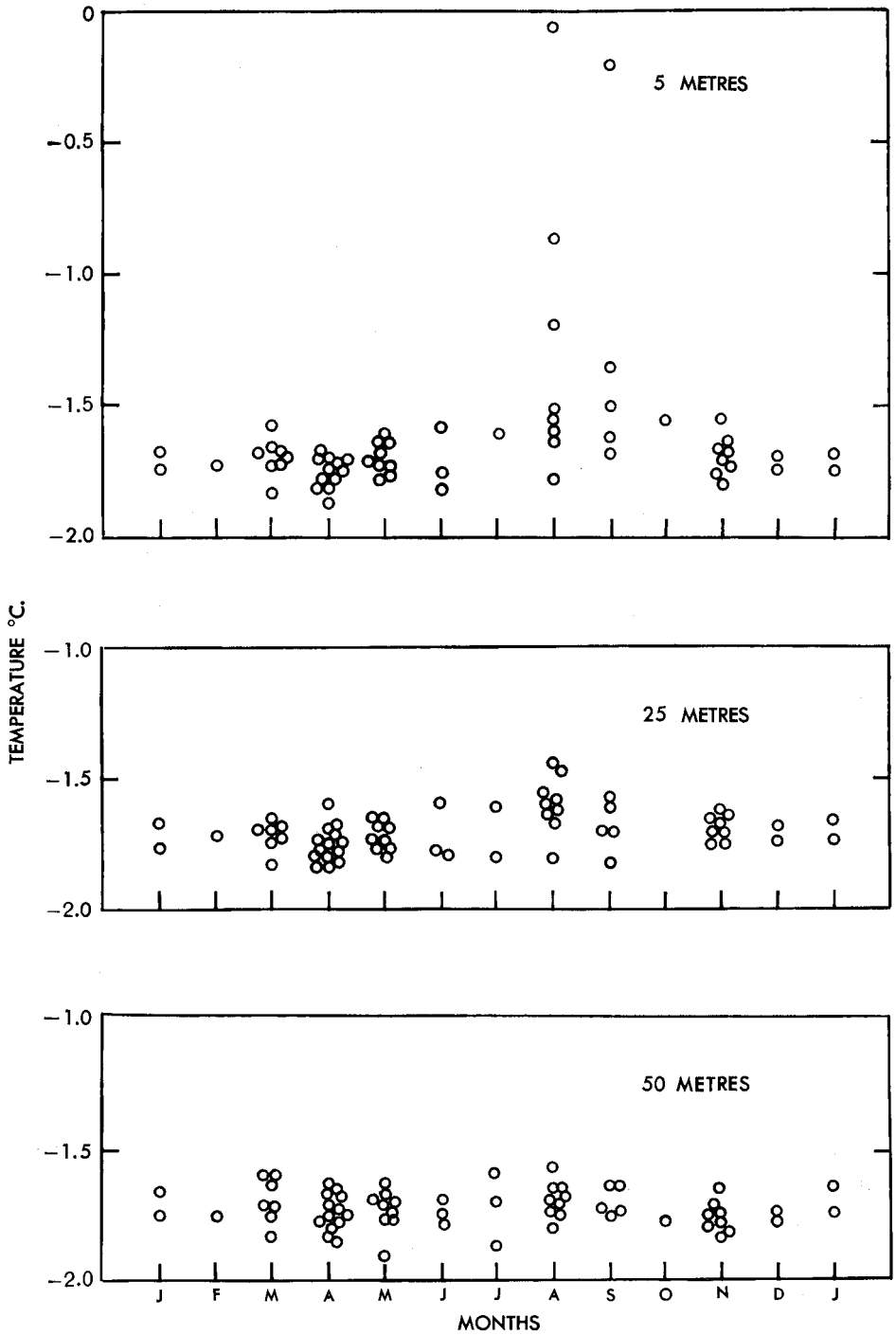
Fig. 4.   Temperatures at the depths of 5, 25, and 50 m., by months.

summertime addition of meltwater produced a slight rise in temperature and decrease in salinity, with a maximum temperature in August, which was nearly as high at 20 m. as at 3 m. but the increase was not noticeable at 40 m.

Shirshov (1944) noted that the surface layer temperatures measured from North Pole-1 were close to freezing in both summer and winter, but that the salt content decreased slightly in summer and increased again the following winter, and that this effect was noticeable only down to 25 m. Treshnikov (1959b), working with data from the *Sedov* and North Pole-2, also observed that seasonally the water temperature fluctuated very little, a few hundredths of a degree C. but the salinity changed significantly by about 2‰.

The temperatures for all stations, grouped by months (Fig. 4), indicate slightly warmer surface water in August and September. This warming may amount to a few tenths of a degree or more at 5 m., but it is barely noticeable, if at all, at 25 m. and below. However, when the freezing point of the water as a function of the salinity (Fig. 5) is taken into account, it is apparent that the water is near its freezing point at all times. The scatter of the temperatures from the freezing points arises, at least in part, from the non-uniformity of the freezing-thawing processes (see above discussion), which nearly obscures the seasonal effects. The seasonal variations may be explained as follows. In the presence of a more or less continuous ice cover, the major seasonal changes are primarily changes in salinity. Any heat that is absorbed goes to melting ice, whether absorbed directly by the ice or first by the water, which then melts the ice. The added meltwater has a higher freezing point than the ambient water, and on being introduced into the surface layer it tends to freeze to the underside of the ice cover, and the liberated heat is re-absorbed by the ice and ambient water. Thus, a rise in temperature of the water due to absorption of heat, up to some stage when the ice cover may become very broken and open, is strictly limited to the increase in the freezing point of the less saline water. In view of the scatter of the temperature values about the freezing point, it is concluded that in the Eurasian Basin the seasonal temperature change is no more than 0.3°C., the difference between the freezing points of the extremes of the observed values of salinity (28-33‰). Normally the change is only about 0.1°C., corresponding to a salinity change of 2‰.

The seasonal change in salinity cannot be adequately interpreted from the available data; such changes can only be properly assessed by following the same water mass. The observations from drifting stations, which tend to follow the near-surface water mass, indicate that the seasonal variation is rather less than the spatial variation within the basin, perhaps only one-half of it.

The penetration of seasonal effects to deeper layers is apparently fairly restricted. In summer, with a decrease in salinity (and hence in density), there is a corresponding increase in the stability of the surface layers, and more energy is required to mix the properties to depth. Estimating from the reports cited and from Fig. 4, noticeable summer seasonal effects reach to
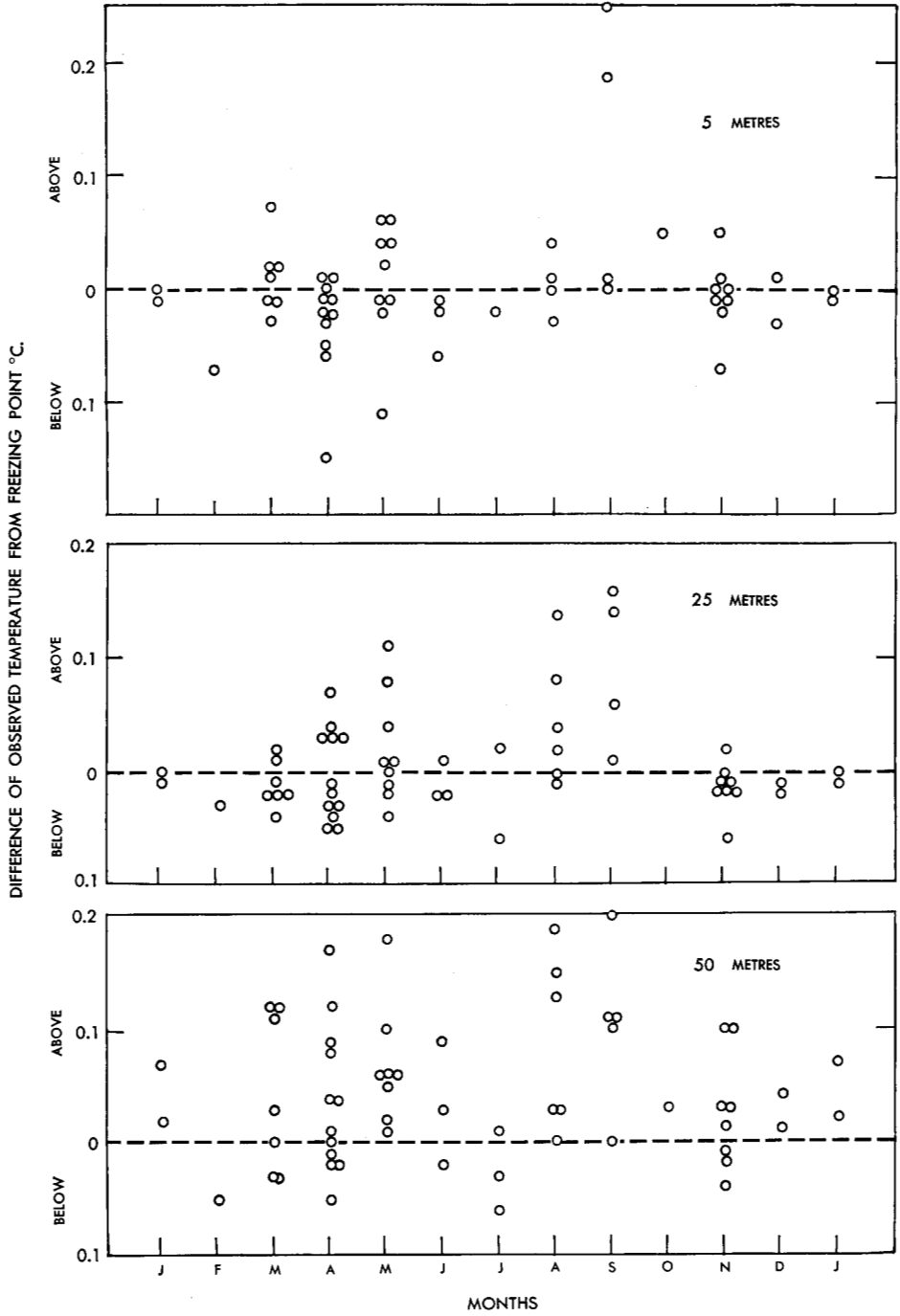
Fig. 5. Deviations of temperatures from freezing point at the depths of 5, 25, and 50 m., by months in °C.

25 m. and perhaps a little deeper, but never to 50 m. In winter, freezing concentrates the water and locally increases its density, leading to an unstable condition and convectional mixing. Because the density is almost wholly salinity controlled in water with these low temperatures, the penetration of surface water by this process may be estimated to reach to that depth where the salinity of the water in the strong halocline matches the winter surface salinity values. Thus, from Fig. 2, the depth of convectional penetration in winter is in general limited to less than 50 m., the sparsity of observations in the vertical precluding a more accurate estimate.

Layer from 25-50 to 100 m. The main characteristics of this layer are a very strong increase in salinity with depth and a temperature that either remains uniformly low or that even decreases slightly, so that a slight minimum may be present in the neighbourhood of 75 m.

Table 3. Seasonal averages of temperature and salinity from 100 m.

Season	Number of observations	Temperature 0°C	Salinity °/oo
Winter-Spring (March, April, May)	29	−1.58	33.85
Summer-Fall (Aug., Sept., Nov.)	22	−1.40	34.11

It is evident that such a vertical distribution cannot result from a direct mixing of the cold, fresher surface layer with the warmer, more saline Atlantic water below, but must be produced and maintained through some advective mechanism. One possibility might be that the salinity gradient is mainly a result of a mixture of surface water with Atlantic water, but in addition water with intermediate values of salinity is produced locally and added to this layer during periods of ice formation. If water in sufficient quantities were produced through this mechanism it would sink to its appropriate density level and thus help maintain a salinity gradient, but it could also reduce the temperature of the water in the layer because it would be initially at the freezing point. However, if this mechanism were effective, there would appear significant seasonal changes in the values of temperature and salinity, not only at the 75- or 100-m. level but throughout the water column above. Variance analyses were applied to all observations from 50 and 100 m. (very few observations are available from 75 m.). Seasonal differences in temperature and salinity significant at the 99.5 per cent level were found at 100 m. (Table 3) but none at 50 m. Therefore the seasonal influences cannot have penetrated to 100 m. locally but must be characteristic of a layer advected from some other part of the Polar Basin.

Two authors have noted that the 25-50- to 100-m. layer must be supplied by advection. Nansen, because he lacked sufficient data, could not be specific about the source of the advected subsurface water but could only theorize
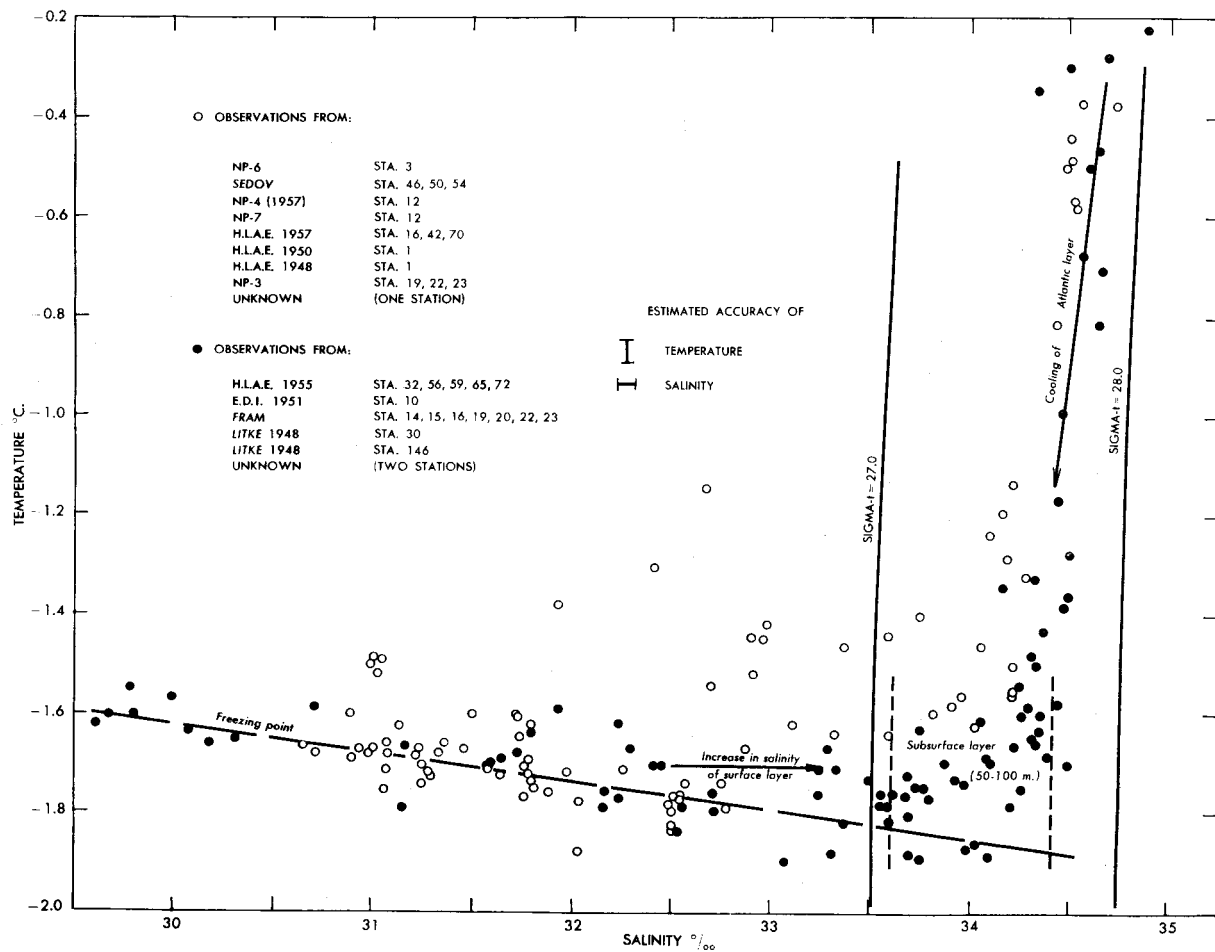
Fig. 6. Temperature-salinity diagram.

that it originated at the surface in some other part of the Arctic Ocean, where the surface water was more saline. He did note, however, that the surface salinities in the bordering seas were sometimes higher than those in the Arctic Basin proper. Treshnikov concluded that this water originated in a band of high-salinity surface water, lying along the continental shelf of Eurasia (Fig. 3), which gradually sinks and spreads to the north under the surface layers of the remainder of the basin.

To determine the characteristics of this particular water and locate its source, the values from the deep-water stations were plotted on a temperature-salinity diagram (Fig. 6) with the station locations shown in Fig. 1b.

The temperature-salinity presentation confirms that the cold but saline water is not a mixture of Atlantic and surface waters, and further there is apparently a geographic variation in the development of the temperature minimum. Water with these characteristics (-1.75 to $-1.90°$C., 33.5 to 34.3‰) appears at subsurface levels at stations adjacent to the Barents, Kara, and Laptev seas but not in other parts of the basin; in the remainder of the basin the extreme low temperatures are absent from this intermediate layer. However, this water is also of such a temperature (-1.2 to $-1.6°$C.) that it cannot be formed by direct mixing of surface and deeper water and hence must be produced and maintained by an appropriate advection.

To locate the primary source of the subsurface water, numerous data from the northern fringes of the bordering seas were examined (Table 1). To use the relative coldness of this intermediate water most efficiently as a tracer, all available data were used to construct a chart of the temperature on the salinity surface S=33.8‰ (Fig. 7), an arbitrary value for the salinity of the core of water that exhibits the low temperature, and a chart of the depth of this salinity surface (Fig. 8). Even though the data are barely sufficient for such representations and the positions of the isolines can only be thought of as very approximate, there appears to be a direct connection between the core of cold water in the Eurasian Basin and the cold water all along the northern fringes of the Barents, Kara, and Laptev seas. Further, this connection is apparently related in some degree to the submarine canyons indenting these seas; this is most noticeable in the northern Kara Sea where one intruding lobe of the $-1.7°$C. isotherm nearly parallels the 200-fathom contour of the St. Anna (Svyataya Anna) Trough. This relationship is not so pronounced elsewhere, e.g., in one area on the northern edge of the Barents Sea east of Northeastland and in three or four areas in the Laptev Sea. In the northeastern part of the latter sea, little judgment may be made about the relationship because there we have neither data nor any accurate knowledge of the topography. Still, there is indication of some connection of the low temperature and specific areas in the northeast Laptev Sea and the northwest East Siberian Sea in the Canadian Basin.

There is also a pronounced slope of this salinity surface. In the central part of the basin it lies at depths of 80-100 m. but along the continental shelf it occupies the 20-40 m. level.

The distribution of properties outlined above leads to the conclusion
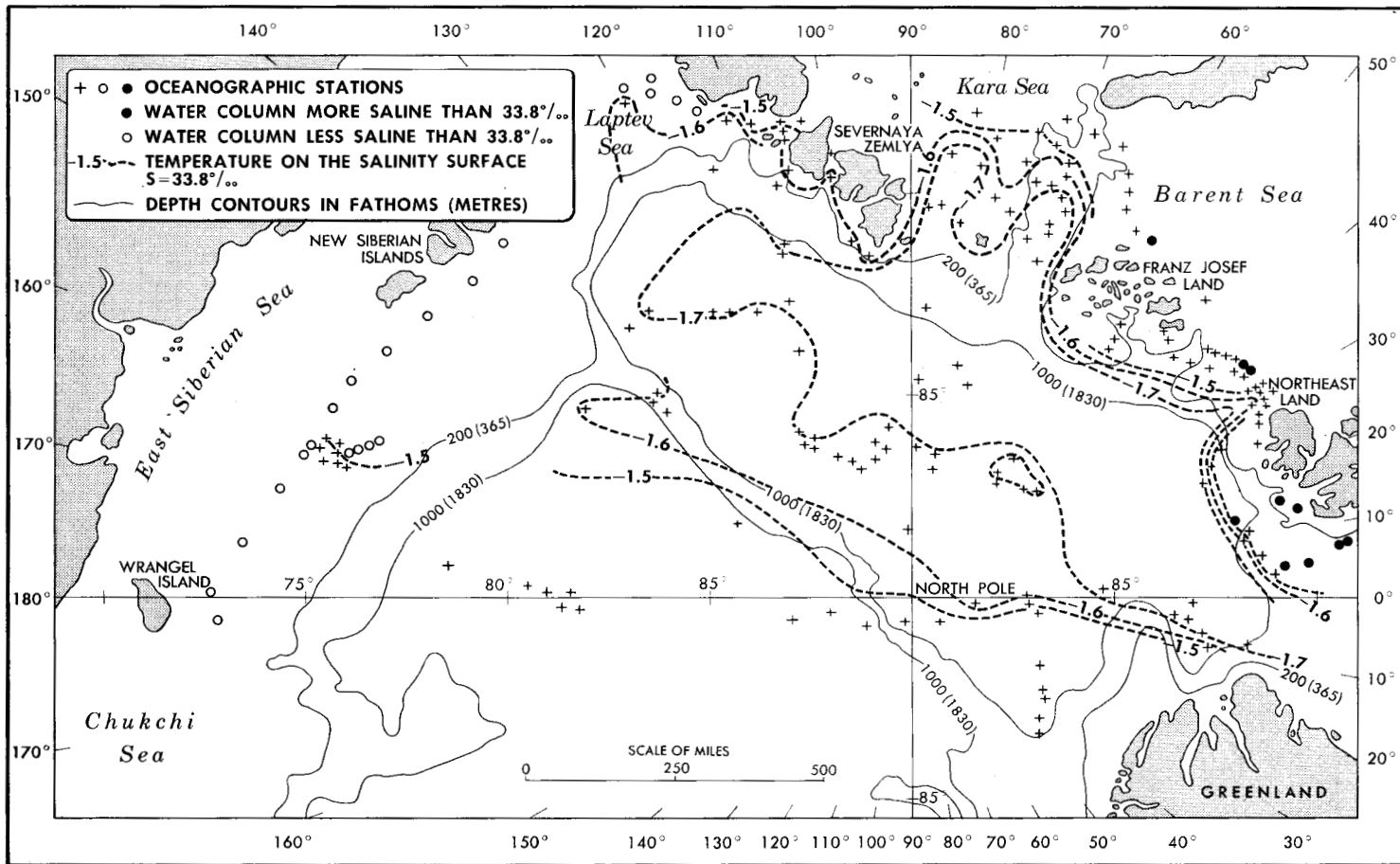
Fig. 7. Temperature on the salinity surface S = 33.8‰.

Legend:

+ o ● OCEANOGRAPHIC STATIONS
● WATER COLUMN MORE SALINE THAN 33.8°/₀₀
o WATER COLUMN LESS SALINE THAN 33.8°/₀₀
-1.5--- TEMPERATURE ON THE SALINITY SURFACE S = 33.8°/₀₀
—— DEPTH CONTOURS IN FATHOMS (METRES)

that the 25-50- to 100-m. layer in the Eurasian Basin is largely a direct result of a subsurface feed of water with the proper temperature and salinity characteristics from along the northern fringes of the peripheral seas and from specific local areas within these seas, and the following model of the circulation is proposed to explain the distribution of properties.

The Barents, Kara, and Laptev seas receive the major part of the continental runoff, which flows north off the continental shelf into the deep-water basin. Underlying the surface along the Eurasian continental slope and in the submarine canyons is water of the highly saline Atlantic layer (Timofeyev 1957). Dynamics similar to those in estuaries develop over the edge of the continental shelf and in the canyons; the surface water moves off the shelf and the deeper water moves toward the shelf and up the canyons where it attains depths considerably shallower than it occupies in the deep-water basin. This highly saline water is here raised to such an extent that it can become mixed with surface water through wind action and through vertical convection that develops in winter and hence is under the influence of environmental cooling and freezing processes. The Atlantic water provides the source of salt to raise the surface salinities from the normal values ($<33‰$) to those of the subsurface minimum (33.8-34‰). This water flows off the shelf and out of the peripheral seas to the north with the surface layer but below it because its density has been increased by the addition of salt. It carries its acquired low temperature characteristic with it, though there would be a seasonal cycle in its temperature; the water producing the slight temperature minimum in the Eurasian Basin would have been produced in winter when the water was close to its freezing point.

The connection of the lowest temperatures of the subsurface layer with the areas of the peripheral seas containing submarine canyons suggests that these canyons play a significant role in the formation of the subsurface layer. This can be shown in the following way. The 34.5‰ salinity surface was chosen to represent the top of the Atlantic layer and a chart of the depth of this surface is presented in Fig. 9. The surface slopes in the manner required by the model; over most of the basin it lies at depths of 140-180 m. but near the edge of the Eurasian continental shelf the surface slopes up, relatively steeply, to depths of less than 60 m. However, in the canyons, notably St. Anna in the northern Kara Sea, the slope of this surface is less steep. Fig. 10 presents two schematic cross sections, derived from Fig. 9, which run from the deep-water basin on to the continental shelf. One section runs directly on to the shelf and the other follows a submarine canyon; the 34.5‰ isoline depicts the slope of the upper surface of the Atlantic water in the two instances. If it is taken that 100 m. is the maximum depth to which mixing processes, e.g., wind action, may reach, then a great deal more area per unit width of the upper surface of Atlantic water comes within reach of the surface processes in the canyons, where the slope of the 34.5‰ surface is more gradual. This can also be seen in Fig. 9 where in St. Anna the area encompassed by the 60- and 100-m. contours is much greater than
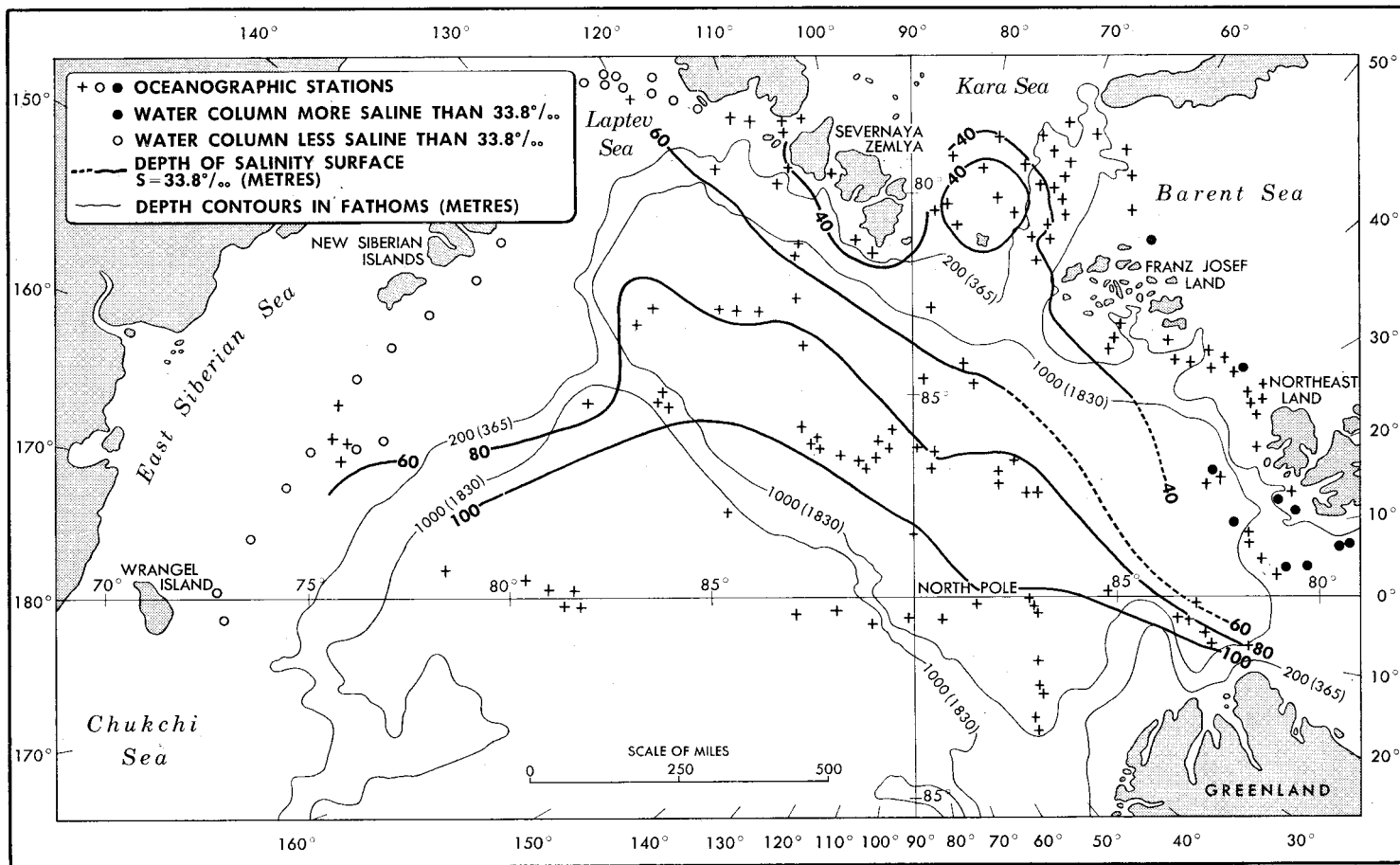
Fig. 8.   Depth of the salinity surface S = 33.8‰.

elsewhere along the edge of the shelf. The wide distribution of a low-temperature subsurface layer over the neighbouring part of the Arctic Ocean indicates a predominance of water of this type, which points to the areas of the submarine canyons as primary sources. Thus, the submarine canyons become important as areas where the largest volumes of the subsurface water are produced.

The model developed here to explain the formation and maintenance of the subsurface layer in the Eurasian Basin is supported by the following arguments:

(1) No indication has been found of any velocity shear within the surface layers of the Arctic Ocean; the few current measurements that have been made (Somov 1954-5) indicate that the surface layers tend to flow as a unit. If this is true then the subsurface layer under discussion would tend to flow outward from the peripheral seas in the same direction as the surface layer, i.e., primarily from the areas of greatest runoff, the Kara and Laptev seas.

(2) The highest values of surface salinities occur along the edge of the Eurasian continental shelf and within the northern parts of the Barents, Kara, and Laptev seas. However, these are the areas where the largest volumes of continental runoff are entrained with the outward-flowing surface water, and hence relatively large amounts of salt are required to give high surface salinities. Thus, here must occur the greatest upward advection of saline water and subsequent mixing with the surface water.

Further, the formation of subsurface feed water is apparently confined to areas along the northern parts of the peripheral seas even though water with the same temperature and salinity characteristics may be formed over much larger areas of the interior Barents Sea and the southern Kara Sea. A close examination of the numerous *Quest* data from the canyon east of Northeastland and the data from the 1931 *Lomonosov* and 1934 *Sedov* cruises over St. Anna indicated that there was a connection of the subsurface type water to the north with the Arctic Ocean, but in the shallow areas along the sides of the canyons and towards the heads of the canyons to the south there was a complete absence of this specific type within the water column. There remains, however, the possibility that some water of the proper type formed in the interior of these seas periodically contributes to the maintenance of the subsurface layer in the Arctic Basin.

The consequences of the particular circulation as described by the model may be summarized thus:

(1) The high temperature of the Atlantic water is a convenient tracer of its movement. Typical temperature and salinity profiles along St. Anna are presented in Fig. 11 and the core of maximum temperature may be traced from about 200 m. at station 7 to 100 m. at station 4 located nearer the head of the canyon. As a first approximation the core may be considered as being maintained by advection and eroded by vertical diffusion. Thus,
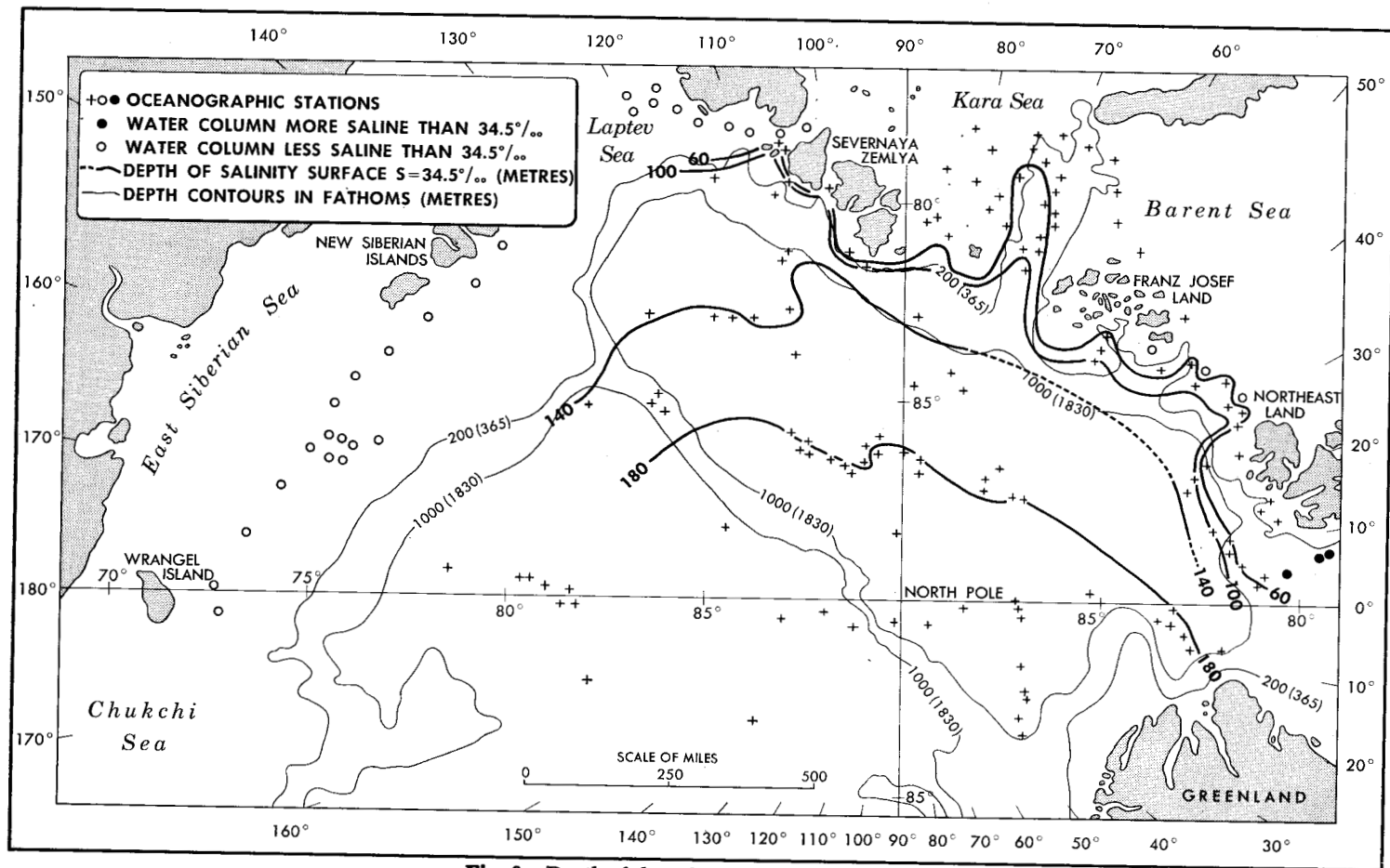
Fig. 9. Depth of the salinity surface S = 34.5‰.

$$U\frac{\partial \theta}{\partial x} = K_z \frac{\partial^2 \theta}{\partial z^2}$$

where U=average velocity along the core, θ=temperature, x and z are distances along the core and in the vertical respectively and K_z=constant vertical eddy coefficient. Solution of this equation between stations 7 and 4 gives K_z/U=1.1 cm. An order of magnitude for the eddy coefficient in this situation might be 1 to 10 cm.²/sec. (Sverdrup et al. 1942), which would indicate velocities of the Atlantic inflow of 0.9 to 9 cm./sec.

(2) The underlying Atlantic water that moves toward the shelf and up the canyons carries with it a large supply of heat; the temperatures of the Atlantic water are everywhere in the basin higher than 0°C. (Timofeyev 1957). The areas where Atlantic water mixes with surface water are areas where this heat is expended with direct influence on the ice cover. For example, an estimate can be made of the upward heat flux in the canyon St. Anna. Upward heat flux is given by

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

where A=area, Q=heat, t=time, ρ=density, and C_p=specific heat. Solution of this equation for the region of the profile of Fig. 11, with a K_z=5 cm.²/sec., gives an upward heat flux at 50 m. of 17×10^{-4} cal. per
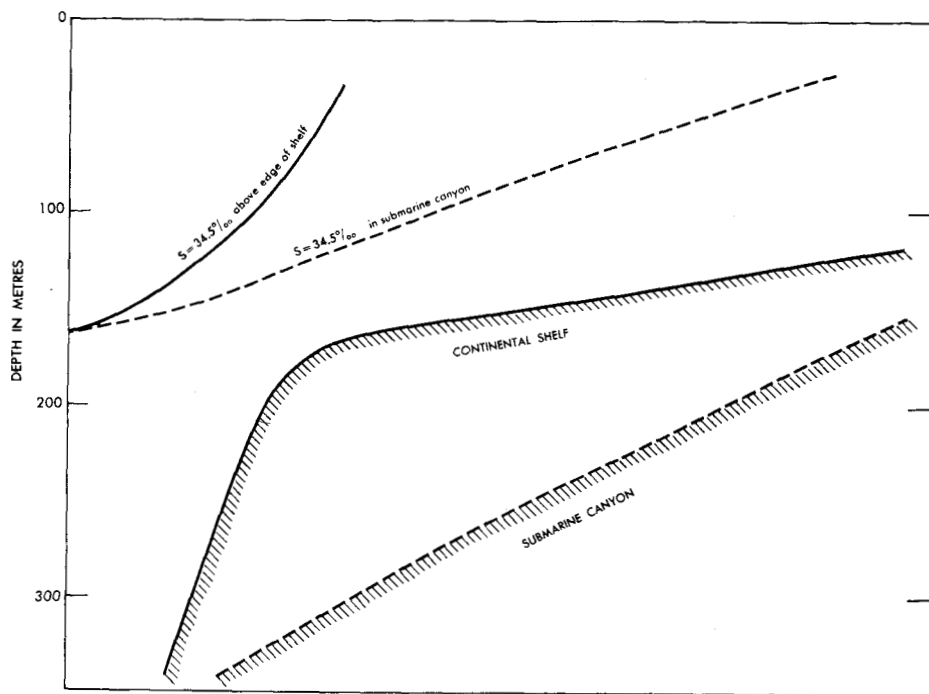

Fig. 10.   Schematic profiles across the edge of the Eurasian continental shelf.
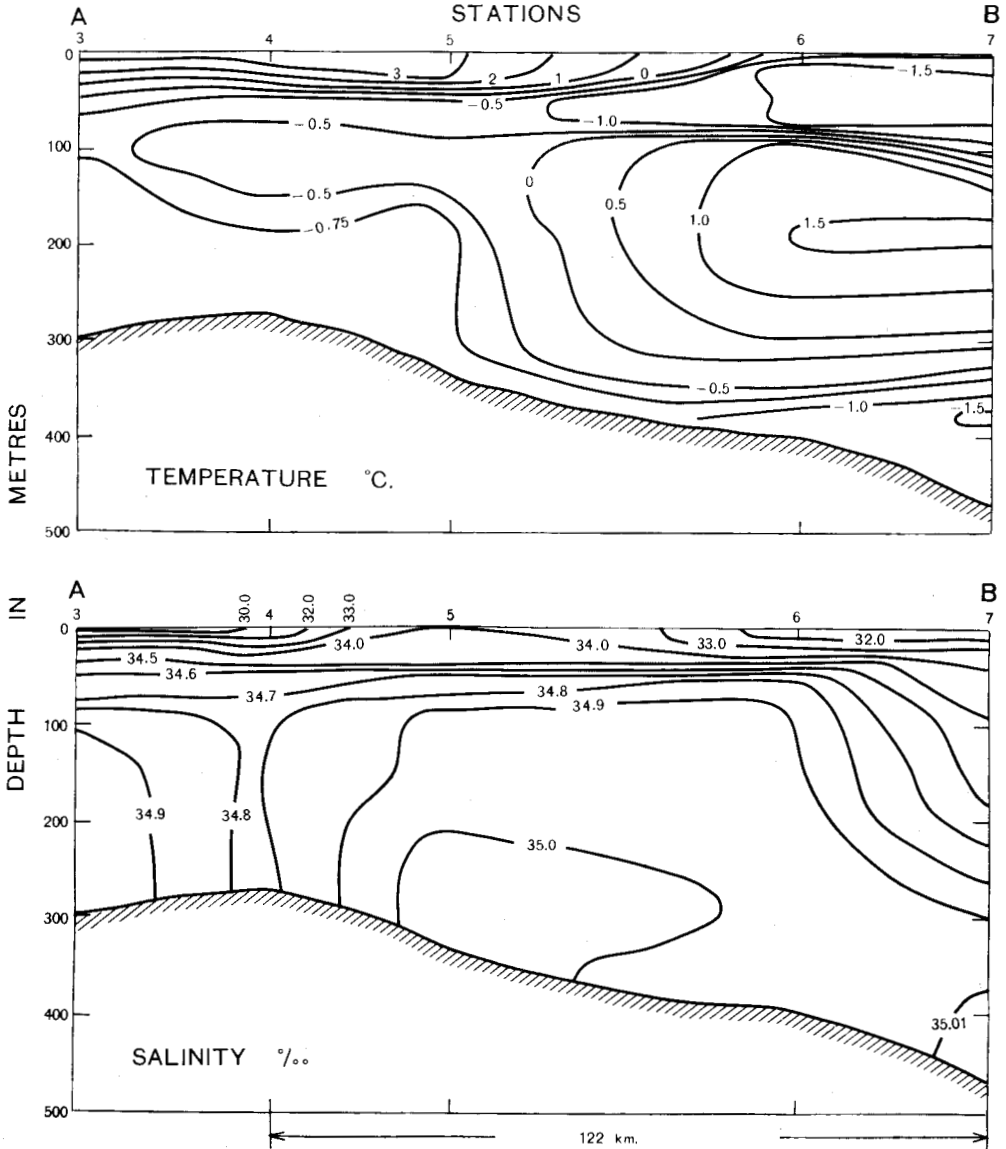
Fig. 11. Profiles of temperature and salinity along the line A — B (St. Anna Trough, Kara Sea) shown in Fig. 1b (after Laktionov 1935).

cm.² per sec. or 147 cal. per cm.² per day. This calculated upward flux of heat is of the same order as the estimated heat loss from the surface in these high latitudes, which is about 330 cal. per cm.² per day gross or 108 cal. per cm.² per day net (Sverdrup *et al.* 1942, p. 99). Hence, this phe-

nomenon could account for the high incidence of polynyas over deep-water parts of the northern Kara Sea and Laptev Sea (Antonov 1958, Kupetskiy 1959, Eskin 1960).

(3) One region traversed by the *Maud* in the north-central East Siberian Sea in the Canadian Basin showed bottom water with salinities as high as 34‰ (this region is found near the 60 m. isobath in Fig. 9). One might speculate that future surveys may disclose a deep-water access to this area from the north and that this is another specific area contributing water in significant quantity to the subsurface layer. But this is only speculation because there is no specific tracer, e.g., low temperature, associated with this water. Its temperature was only —1.4°C., due apparently to the mechanics of bottom water formation on the East Siberian shelf (cf. Sverdrup 1929) and thus this water does not contribute to the cold subsurface layer but could contribute locally to the subsurface layer to the north of the East Siberian Sea in the Canadian Basin, which exhibits temperatures only slightly warmer than —1.4°C.

Summary and Conclusions

(1) Surface water in the Eurasian Basin of the Arctic Ocean occupies approximately the upper 200 m. It is more or less continuously supplied by continental runoff from Siberia that, in traversing the peripheral Arctic seas, the Barents, Kara, Laptev, and East Siberian, mixes with and entrains a few hundred times its volume of saline water. The surface water then tends to flow on the most direct route to the exit from the basin between Spitsbergen and Greenland. The surface water can be divided into three layers on the basis of temperature and salinity characteristics.

(2) The surface layer, 0 to 25-50 m. has a broad range of salinity, from about 28 to 33.5‰, depending on geographic location in the basin; the water nearer the bordering seas tends to have a lower salt content than closer to the basin exit. Surface-layer temperatures are always close to freezing, for the salinity of the water, due to the continuous presence of an ice cover. There is, however, some deviation from these temperatures because the ice cover is not uniform, leading in turn to non-uniformities in the action of freezing-thawing processes. This layer down to below 25., but rarely to 50 m., is subject to seasonal variation, being less saline and slightly warmer in August and September than during the remainder of the year. These effects are greater near the surface and less at 25 m.; the seasonal variation in salinity is of the order of 2‰ and the temperature change 0.1°C. Vertical convection induced by the freezing of ice is limited to this layer by the strong density increase immediately below.

(3) The subsurface layer, 25-50 to 100 m., contains a strong halocline such that the salinity at 100 m. is close to 34‰ or greater, whereas the temperature changes little. In the area paralleling the continental slope from Spitsbergen east to the Laptev Sea this layer remains nearly as cold as the surface layer, and the temperature may even decrease slightly to

a minimum at 60-80 m., but in other areas the temperature may rise a few tenths of a degree above the freezing point. This layer also may be supplied by advection from the northern fringes of the peripheral seas, but in large part from specific local areas. The water of this layer is intimately connected with the submarine canyons that indent the bottom of the seas bordering the Arctic Basin.

(4) The layer from approximately 100 m. down to the Atlantic layer, the core of which lies at 250-300 m. in the Eurasian Basin, has properties intermediate between those of the subsurface layer and the Atlantic water. It results from mixing of the two layers, and because the surface layer, the subsurface layer, and the Atlantic layer are more or less continuously replenished there is a quasi-steady-state distribution of properties at this level.

(5) A model is proposed to explain the dominant distribution of the cold subsurface layer in the Eurasian Basin. The submarine canyons along the continental slope are, in effect, sunken estuaries providing egress locally into the peripheral seas. Highly saline Atlantic water moves along the Eurasian continental slope and enters the canyons. Surface water flows north out of the canyon and away from the continental slope, and, through dynamics similar to those of estuaries, the Atlantic layer here rises to shallower depths. At the shallower depths the Atlantic water is mixed with surface water and hence comes under the influence of climatic conditions; the temperature of this mixture, still relatively saline, is reduced to its freezing point and then the mixture feeds back into the Eurasian Basin as the lower part of the subsurface layer. The submarine canyons thus provide a large area in which mixing and cooling take place, and as such are important as primary sources of the subsurface water.

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