

On the halocline of the Arctic Ocean*

KNUT AAGAARD†, L. K. COACHMAN† and EDDY CARMACK‡

(Received 13 November 1980; accepted 17 December 1980; final revision received 18 December 1980)

Abstract—The cold upper halocline of the Arctic Ocean is maintained by large-scale lateral advection from the adjoining continental shelves, where dense and saline shelf water is produced during freezing; the salinization of the water column is especially pronounced in certain areas of persistent ice divergence. Estimates show the annual rate at which the dense shelf water feeds into the Polar Basin is probably in the neighborhood of $2.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$; this is of the same order as the inflow of warm and saline water from the Atlantic. A consequence of this process is that the halocline must be a heat sink for the underlying Atlantic water, thereby shielding the ice cover from an upward heat flux. The Atlantic water is thus linked rather directly to the enormous shelf seas that border the Polar Basin. Proposed massive river diversions in the Arctic could, by increasing the shelf salinities and driving a deeper flow into the interior, cause a thinning of the halocline and place the Atlantic water in more direct contact with the surface mixed layer.

INTRODUCTION

THE TWO major vertical hydrographic structures of the Arctic Ocean are a pronounced cold halocline between 50 and 200 m and a thick layer of warm water derived from the Atlantic Ocean and centered near 300 to 500 m (see e.g., Fig. 2 in TRESHNIKOV and BARANOV, 1972; or Figs 4 and 6 in COACHMAN and AAGAARD, 1974). Because of the strong salinity dependence of density at low temperatures, the halocline is also a marked pycnocline. The significance of this pycnocline is that it insulates the large sub-surface store of sensible heat from the surface ice cover and the atmosphere. An important feature of the geography is that about one-third of the Arctic Ocean area is shelf seas (Fig. 1).

The low salinity of the surface layer is due in part to continental runoff, primarily via the Siberian rivers but also from other rivers and from the North Pacific through Bering Strait. However, the seasonal melting and freezing of sea ice may also be important in maintaining the low surface salinity. For example, AAGAARD and COACHMAN (1975) estimated that the total fresh water addition from runoff and the Bering Strait inflow is about $0.15 \times 10^6 \text{ m}^3 \text{ s}^{-1}$. If distributed over the $8 \times 10^6 \text{ km}^2$ of the deep Polar Basin, this represents, annually, a column of fresh water 60 cm high. A comparable thickness of ice, with a bulk salinity of perhaps 5,§ melts in the basin each summer (THORNDIKE, ROTHROCK, MAYKUT and COLONY, 1975), adding its relatively fresh water to the surface. The ensuing stratification inhibits mixing of the upper layer, while salt separation during freezing of the

* Contribution No. 1187 from the Department of Oceanography, University of Washington, Seattle, WA 98195, U.S.A.

† Department of Oceanography WB-10, University of Washington, Seattle, WA 98195, U.S.A.

‡ Department of the Environment, N.W.R.I. Branch, 4160 Marine Drive, West Vancouver, B.C. V7V 1N6, Canada.

§ In this paper no units of salinity are used. The numbers given are salinity $\times 10^3$, Editor.



Fig. 1. Geography and partial bathymetry of the Arctic Ocean, showing the three stations in Figs 3 and 4 and the area covered by Figs 8 to 12. The exchange of Atlantic water is indicated schematically by the arrows.

following winter drives convection and, if the convection is penetrative, produces a deeper halocline. Under these circumstances the melting-freezing cycle would produce a low-salinity surface layer, independent of external fresh water sources.

Our purpose in the present paper is to examine the maintenance of the halocline, a layer of cold water 3 to 4 times as thick as the upper low-salinity layer. We shall show that its properties can only be explained through lateral advection, that the source of the advected water must be the shelf seas during winter, and that the quantity of water produced annually is of the same order as the inflow of Atlantic water. A diagram of the temperature and salinity structure of the Arctic Ocean and the proposed mechanism of halocline maintenance is shown in Fig. 2.

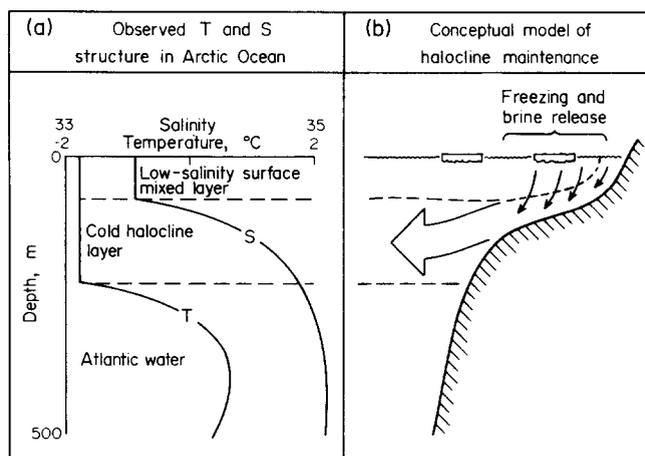


Fig. 2. Schematic representation of the temperature and salinity structure in the upper Arctic Ocean and its maintenance.

THE MAINTENANCE OF THE HALOCLINE

Figure 3 shows the observed upper temperature and salinity structure at three winter stations from widely separated parts of the Arctic Ocean. In each case the temperature remains close to the freezing point to depths about twice that of the isohaline upper layer. This extension of the isothermal surface layer into the halocline was first noted by NANSEN (1902) and it has since been discussed by SHIRSHOV (1944), TIMOFYEV (1951, 1960), TRESHNIKOV (1959), COACHMAN and BARNES (1962), and others. The feature appears to be most pronounced in the Eurasian Basin, for in much of the Canadian Basin the thermal structure is complicated by the presence of a temperature maximum near 75 to 100 m and representing a summer contribution of water from the Bering Sea (contrast COACHMAN and BARNES, 1962, Fig. 2 with KINNEY, ARHELGER and BURRELL, 1970, Fig. 8). Except for SHIRSHOV and TIMOFYEV, who believed that the cold upper part of the halocline resulted from a mixture of surface and Atlantic waters, there has been a general agreement that the upper halocline layer must be supplied by advection. The crux of the argument can be seen from the temperature-salinity (T-S) diagram (Fig. 4) for the three stations shown in Fig. 3. Linear mixing of waters from the surface and Atlantic layers cannot produce the cold water of the upper halocline. We must therefore look to direct injection of this water, which is near the freezing point and 34 or more in salinity. Furthermore, because salinities as high as those in the upper halocline are never found close to the surface in the central Arctic (COACHMAN and BARNES, 1962), a mixed-layer source there is unlikely (in contrast to the formation of 18° mode water in the North Atlantic, for example). Rather, the source must be somewhere on the periphery. TRESHNIKOV (1959) proposed that it lies along the continental slope from Spitsbergen to the Laptev Sea and COACHMAN and BARNES (1962) proposed the submarine canyons indenting the shelves in this same area to be primary sources.

Additional evidence that water in the upper part of the halocline is maintained by lateral advection is the distribution of nutrients. Winter profiles of reactive silicate, for example, typically show a strong vertical gradient coincident with that of salinity, but above the

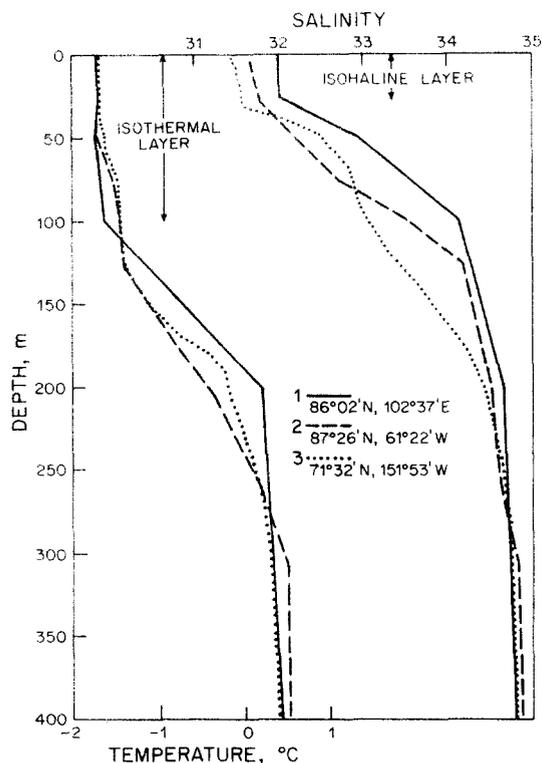


Fig. 3. Vertical temperature and salinity distribution in the upper 400 m as observed during winter at the three locations indicated in Fig. 1. Station 1 was occupied from 16 to 17 February, 1956 (TIMOFEYEV, 1960), Sta. 2 on 20 February, 1964 (TRIPP and KUSUNOKI, 1967), and Sta. 3 on 11 March, 1977 (Sta. W27-24 from AAGAARD, 1978).

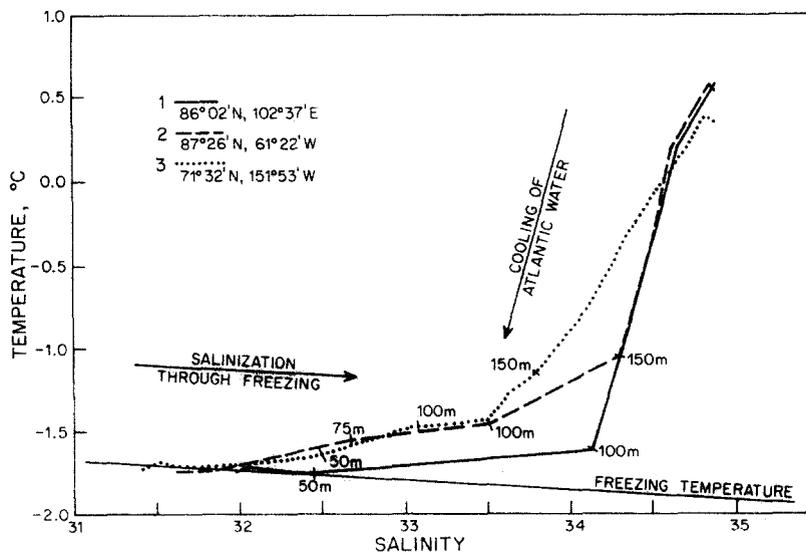


Fig. 4. Temperature-salinity diagram for the stations shown in Fig. 3. Two possible mechanisms for producing cold halocline water are indicated by the arrows.

temperature gradient, i.e., at depths where the temperature is still within a few tenths of a degree of freezing (R. MOORE and M. LOWINGS, personal communication). Similarly, at the station at $87^{\circ}26'N$, $61^{\circ}22'W$ (Fig. 3) the reactive phosphate concentration more than doubled from 51 to 76 m, whereas the temperature was still within $0.2^{\circ}C$ of freezing (data from TRIPP and KUSUNOKI, 1967).

We can estimate the production rate of the cold source water in the following way. Figure 4 shows that the discontinuity in the slope of the T-S correlation occurs near 150 m. As the upper isopycnal layer does not extend below about 50 m, the cold upper part of the pycnocline (above the T-S slope discontinuity) has a typical thickness of 100 m. It is this layer that Fig. 4 suggests is mixed linearly with the core of Atlantic water. The fresh water residence time for the Arctic Ocean has been calculated to be about 10 years (AAGAARD and COACHMAN, 1975) and if the upper pycnocline water is assumed to have a similar time scale, then the production rate of this water is $100\text{ m} \times 8 \times 10^6\text{ km}^2 \div 10\text{ y} = 2.5 \times 10^6\text{ m}^3\text{ s}^{-1}$. Figure 4 suggests that the more saline part of this water has characteristic salinities in the range 33.5 to 34.5.

The production rate of cold source water can also be estimated by considering the transformation within the Polar Basin of the Atlantic water (Fig. 5). When this water first enters the basin west of Spitsbergen, its mean temperature and salinity are near $2.2^{\circ}C$ and 35.0, but east of Greenland these properties have decreased to about $0.5^{\circ}C$ and 34.9 (AAGAARD and GREISMAN, 1975). If we assume that this cooling and freshening is due to linear mixing with overlying water near the freezing point, as suggested in Fig. 4, then the latter must have a salinity of about 34.7, which is much higher than is observed in the surface layer in winter in the central Arctic Ocean. The mixing ratio is close to 60% of inflowing Atlantic water and 40% of the cold, saline type (Fig. 5). Recent estimates of the mean annual rate of inflow of Atlantic water, based on direct current measurements, have ranged from more than $7 \times 10^6\text{ m}^3\text{ s}^{-1}$ (AAGAARD and GREISMAN, 1975) to less than $4 \times 10^6\text{ m}^3\text{ s}^{-1}$ (unpublished observations). Linear mixing as described in Fig. 5 would then require a production rate of the cold, saline source water (type B) in the range 2.5 to

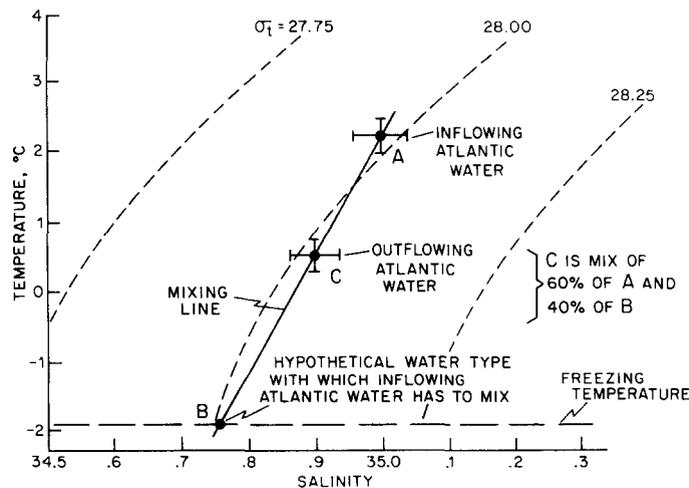


Fig. 5. Hypothetical mixing diagram for Atlantic water in the Arctic Ocean.

$5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$. The above salinity and production rate estimates for this water are upper bounds because Atlantic water probably also mixes with underlying deep water and because part of the upward salt flux may occur by double diffusion.

UPWELLING AND COOLING OF ATLANTIC WATER

Two possible mechanisms that can produce source water of the required temperature and salinity are the cooling of Atlantic water and the salinization of surface waters by brine expulsion during freezing (Fig. 4).

The first mechanism requires that the saline Atlantic water be raised from its relatively deep position in the basin to lesser depths, where it is cooled and freshened. This was proposed by COACHMAN and BARNES (1962) to occur in the canyons of the northern Kara Sea. Indications of upwelling were also observed in the northern East Siberian Sea by SVERDRUP (1929) and in Barrow Canyon by MOUNTAIN, COACHMAN and AAGAARD (1976). Recently we have seen direct evidence for upwelling on the northern Alaskan shelf. For example, Fig. 6 shows two early winter sections across the slope and outer shelf of the Alaskan Beaufort Sea. The warm ($T \sim 0^\circ\text{C}$) and saline ($S > 34.5$) water along the shelf break at 100 to 120 m can only have come from depths in the Arctic Ocean greater than

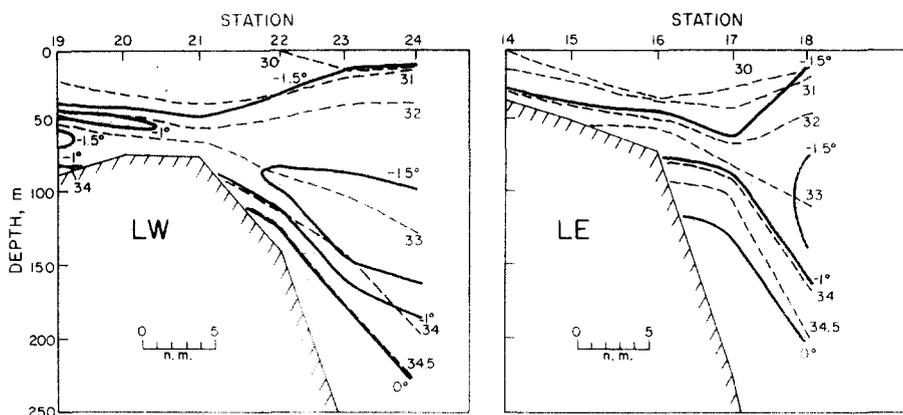


Fig. 6. Two sections of temperature and salinity across the Beaufort Sea shelf during November, 1976. Section locations are shown in Fig. 12. Adapted from AAGAARD (1977).

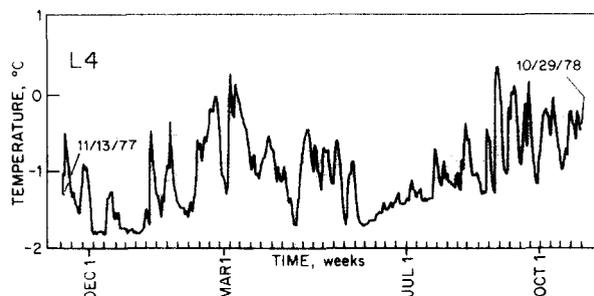


Fig. 7. Daily mean temperature 15 m above the bottom during 1977 and 1978 near the Beaufort Sea shelf break at the location shown in Fig. 12. The water depth was 192 m. Adapted from AAGAARD (1979).

about 220 m. That upwelling is common on this shelf is indicated by the temperature record 15 m above the bottom for a one-year period (Fig. 7). Each pulse of elevated temperature represents the presence of upwelled water. Although it is clear that saline water characteristic of the upper part of the Atlantic layer may be elevated onto the shelves, where cooling and mixing might in winter change its properties to those of the upper pycnocline water, the rates and the mechanisms of the transformation remain uncertain. Our conclusion is that some contribution to the pycnocline from this source is probable, but its overall importance is unknown.

THE SALINIZATION OF SHELF WATER

The second possible mechanism for the production of cold halocline water is salinization of shelf water during winter. Sea ice contains a variable amount of salt, depending primarily on growth rate, with an average value of about 5 (MALMGREN, 1927). Assuming that the expelled salt is added uniformly to the underlying water, a simple continuity calculation yields the thickness of new ice required to raise the initial salinity of a given column of water to a specified salinity. The higher the initial salinity and the shallower the water, the more effective this mechanism is in creating a dense water type at the freezing point, a situation similar to that in which thermal bars are formed in lakes (HUANG, 1972).

To assess the potential of the above mechanism, we have surveyed the data available from the various arctic shelves. For each station we calculated the growth of ice required to raise the water column to a given salinity, the value of which corresponds to the break in the T-S diagram at the closest stations within the Polar Basin. For the region from 20 to 50°E this was 34.5, for 60 to 100°E 34.0, and for the remaining shelves 33.5. In this way we model regional differences in the salinity of the halocline. The calculation was extended only to 50 m (or the bottom in water depths less than 50 m) because this corresponds to the observed depth of the upper part of the halocline in the Polar Basin. The calculated ice growth is probably the *maximum* required for forming halocline water because most station data are from summer when the fresh water accumulation is maximal and, in general, some of the fresh water leaves the shelf prior to ice formation. A point of reference is that typically about 2 m of ice is grown each winter over the arctic shelves (see e.g., HIBLER, 1979).

The region from Spitsbergen to Franz Josef Land (Fig. 8) is a particularly likely source of halocline water because relatively little ice growth is required anywhere along this entire shelf edge to produce sufficiently saline water. In the Kara Sea (Fig. 9) a likely source is between Franz Josef Land and Novaya Zemlya, where there is a saline inflow from the Barents Sea and probably also through Svataya Anna Canyon. A second possible source area appears over the Voronin Canyon west of Severnaya Zemlya. In the Laptev Sea (Fig. 10) the summer salinities are in general far too low to allow winter production of water sufficiently saline to feed the halocline. However, there are no data from the outermost part of the shelf, while Fig. 11 suggests that this northern area might be a source. There is also some potential close to shore, where the water is sufficiently shallow. A similar effect can be seen in the southeastern Kara Sea (Fig. 9).

All of the above sources from Spitsbergen to the New Siberian Islands would primarily feed the Eurasian Basin halocline. Primary sources for the Canadian Basin must be sought farther east. Calculations for the East Siberian Sea (not shown) indicate a required ice growth in excess of 4 m, so that this sea is an unlikely source. However, the East Siberian

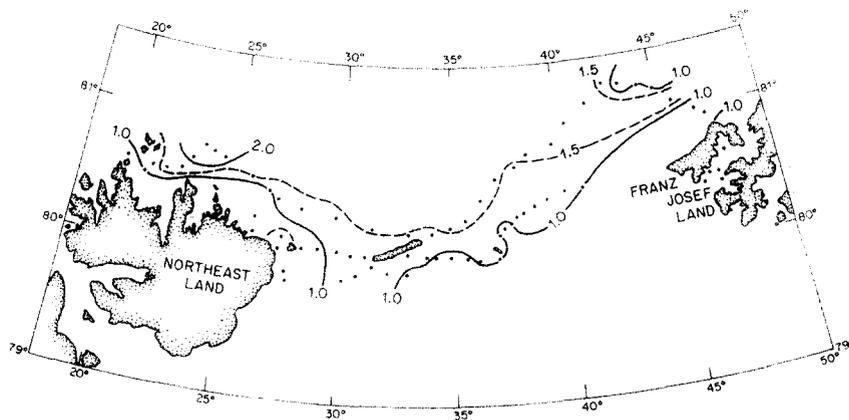


Fig. 8. Ice growth (m) required to raise the salinity of the summer water column on the shelf to 34.5. This salinity corresponds to the break in the T-S diagram at the closest off-shelf stations. The area location is shown in Fig. 1.

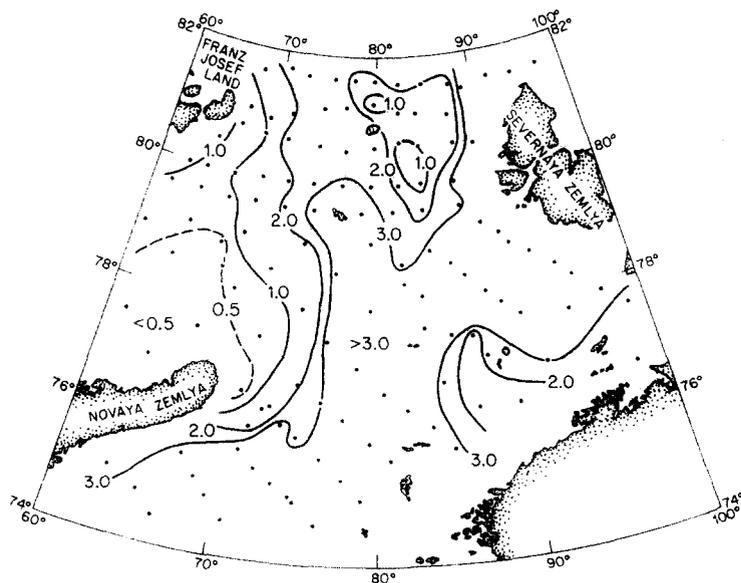


Fig. 9. Ice growth (m) required to raise the salinity of the column on the shelf to 34.0. This salinity corresponds to the break in the T-S diagram at the closest off-shelf stations. The area location is shown in Fig. 1.

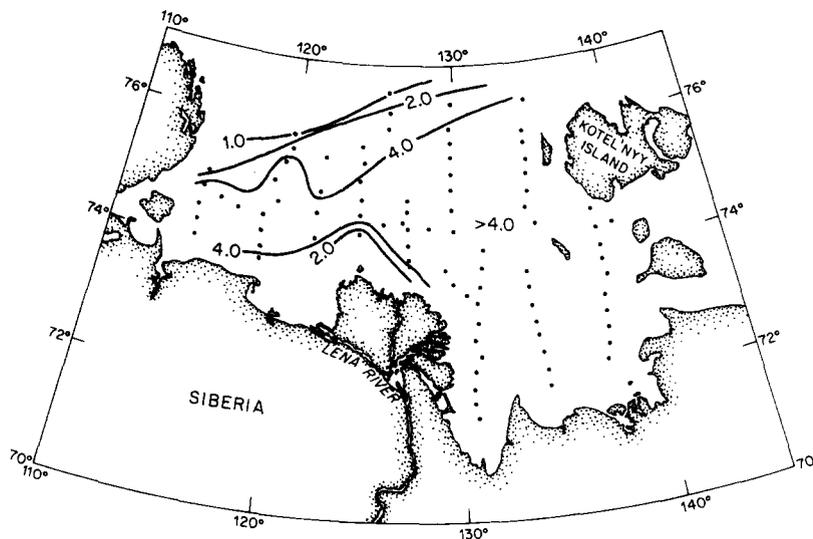


Fig. 10. Ice growth (m) required to raise the salinity of the summer water column on the shelf to 33.5. This salinity corresponds to the break in the T-S diagram at the closest off-shelf stations. The area location is shown in Fig. 1.

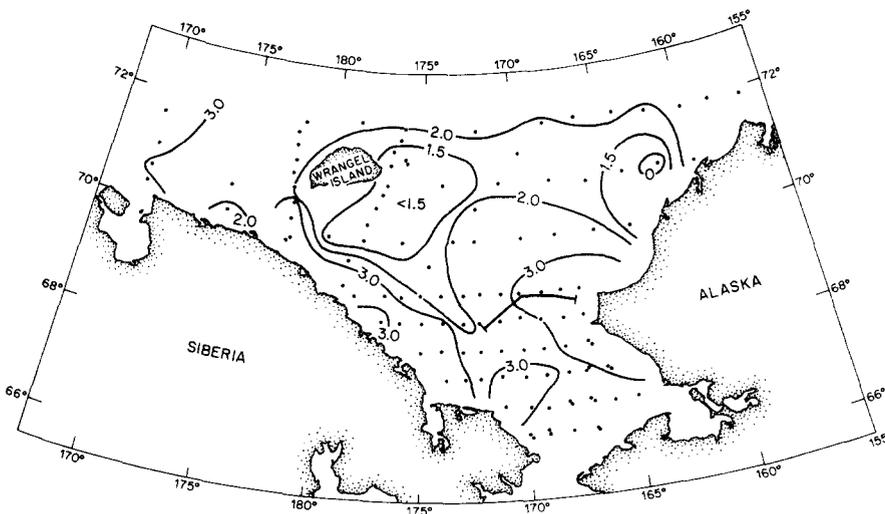


Fig. 11. Ice growth (m) required to raise the salinity of the summer water column on the shelf to 33.5. The salinity corresponds to the break in the T-S diagram at the closest off-shelf stations. The area location is shown in Fig. 1. The location of the section shown in Fig. 14 is indicated by the heavy bracketed line near 69°N.

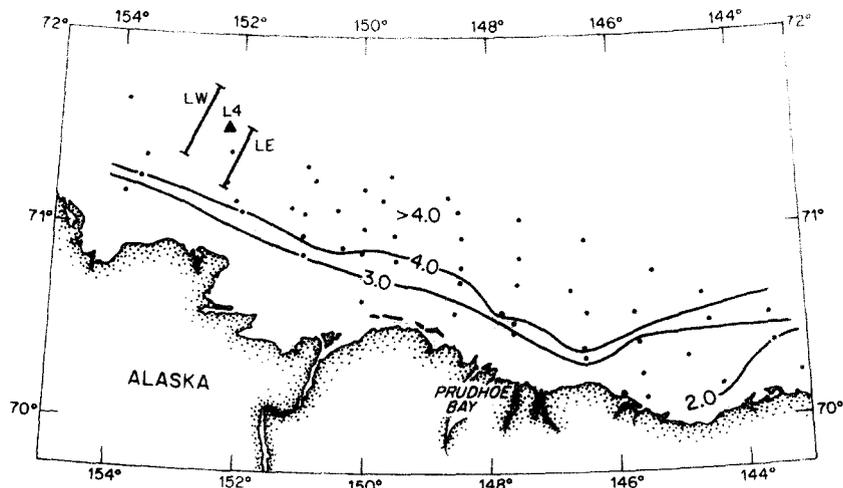


Fig. 12. Ice growth (m) required to raise the salinity of the summer water column on the shelf to 33.5. This salinity corresponds to the break in the T-S diagram at the closest off-shelf stations. The area location is shown in Fig. 1. The location of the sections shown in Fig. 6 is indicated by lines LW and LE, and the site of the temperature record shown in Fig. 7 is indicated by triangle L4.

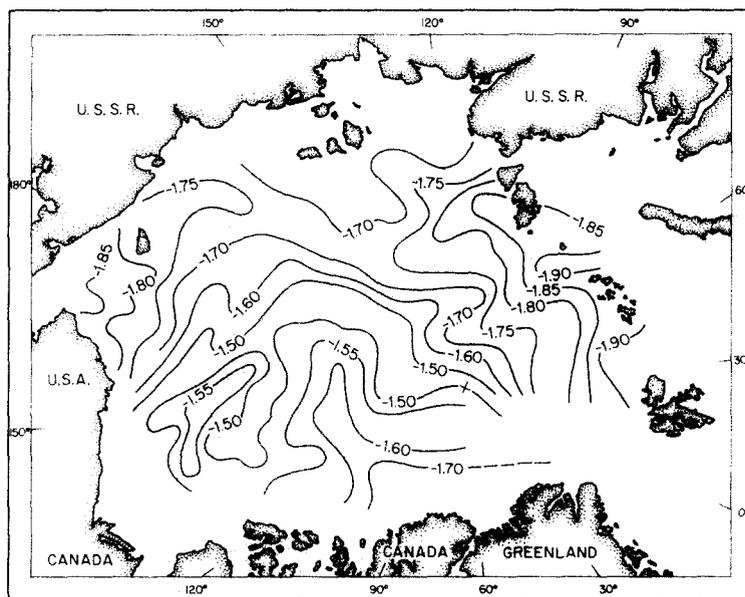


Fig. 13. Minimum temperatures in the halocline of the Arctic Ocean (adapted from TRESHNIKOV and BARANOV, 1972).

Sea shelf is very broad (up to 900 km) and there are no oceanographic observations available from the outer part of the shelf, so our conclusion is tentative. However, the Chukchi Sea (Fig. 11) shows some potential and we shall consider the area later in detail. Continuing eastward, the narrow Beaufort Sea shelf (Fig. 12) seems capable of producing saline water at the freezing point only in shallow water very near shore. This agrees with the observations of WISEMAN (1977), who found salinities of 35 restricted to lagoons inside the barrier islands. In the Canadian Beaufort Sea the potential for production of saline water appears to be even less than in the East Siberian Sea because of the low salinities. However, a possible exception is Amundsen Gulf between Banks Island and the mainland.

No data are available from the shelf areas along the northern fringes of the Canadian Arctic Archipelago. However, the distribution of minimum temperatures in the halocline of the Polar Basin (Fig. 13, adapted from TRESHNIKOV and BARANOV, 1972) shows a cold tongue north of the archipelago, suggesting a regional contribution.

WINTER CONDITIONS IN THE CHUKCHI AND BERING SEAS

We noted above that the summer data suggest a potential for the winter production of saline water in the Chukchi Sea. In February and March, 1977 we occupied an east-west CTD (conductivity-temperature-depth) section across the central Chukchi by helicopter, the only winter section from the region. The temperature was everywhere within a few hundredths °C of the freezing point; the salinity distribution is shown in Fig. 14. In general the salinities are remarkably high, typically 2 to 4 higher than in summer. Water more saline than about 34 had previously only been found in the Chukchi in isolated lagoons, whereas in our data the mean near-bottom salinity was greater than 34.2 over the entire 250-km long section. The highest salinities were in the eastern Chukchi, with Sta. 33 having a mean salinity in excess of 34.7. All of these stations are clearly capable of supplying cold, saline water to the deep parts of the pycnocline.

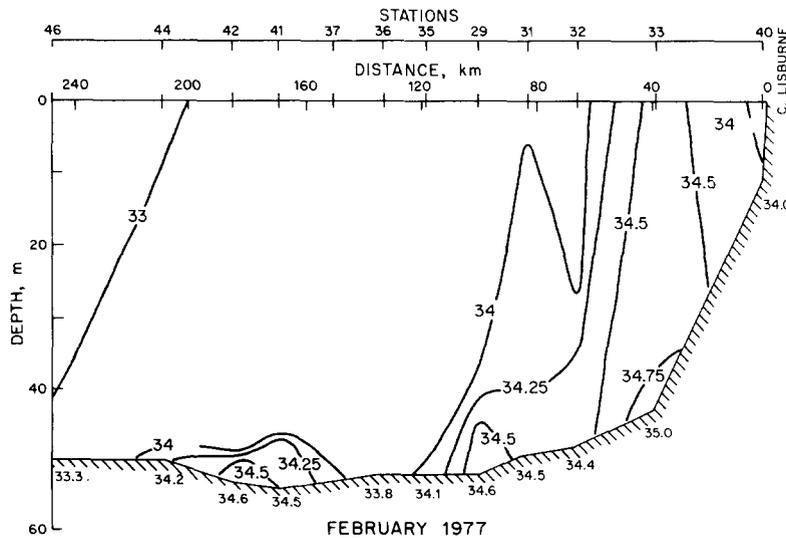


Fig. 14. Salinity section across the Chukchi Sea, February to March, 1977. The section location is shown in Fig. 11.

A typical mean salinity in summer for the section in Fig. 14 would be about 32.5 (COACHMAN, AAGAARD and TRIPP, 1975) and to increase its salinity to 34 would require the formation of over 2 m of ice. For Sta. 33 alone, a change from 32 in summer to 34.7 in late winter would require that about 4 m of ice be formed. To account for these large amounts of ice growth, we hypothesize that the northern Bering-Chukchi Sea region contains numerous areas of local surface divergence where the ice cover is maintained thin and broken. As the rate of ice growth decreases exponentially with ice thickness (Stefan's equation, e.g., NEUMANN and PIERSON, 1966, p. 85), the effectiveness of thin ice in salinization of the water column is considerable. For example, with an air temperature of -10°C , about 8 days are required to form 30 cm of ice. On the other hand, if the newly formed ice is continually removed, the equivalent ice growth over 8 days would be close to 80 cm. It is precisely this mechanism of mechanical removal of ice that GILL (1973) proposed is important in the production of dense shelf water in the Antarctic.

In support of the above hypothesis, the highest salinities in Fig. 14 coincided with a large region of persistent open water or thin and broken ice in the eastern Chukchi. Satellite images (Fig. 15) show that open or nearly open water in winter is a frequent occurrence in much of the Bering Strait region, notably along the shores southeast from Point Hope and Bering Strait, along the north shores of Norton Sound and the Gulf of Anadyr, and south of St Lawrence Island. As all the water on the shelf was at freezing point, only mechanical removal of the ice could maintain an open area. Such removal was almost certainly accomplished by surface divergence due to prevailing northeasterly offshore winds (compare CARLETON, 1980). Two other areas of thin ice can be seen in Fig. 15—in the Strait of Anadyr and in Bering Strait, both in the lee of arch-shaped dams of ice in these two narrow straits. All of these areas are thus focal points for effective salinization of the water column in winter.

Where circulation and flushing are weak, saline water can persist into summer. For example, near-bottom salinities in Norton Sound during July, 1977 (Fig. 16) show a band of inordinately high salinities off the north shore, corresponding closely to the location of open water in winter. On the other hand, during the previous summer (1976) near-bottom salinities were less than about 30, so that the dense water may not be formed every winter.

Rough calculations of the amount of water more saline than 34 that might reasonably be produced in the Chukchi each winter are of order $0.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ (annual average). Because of the prevailing northward flow through Bering Strait, this production would be supplemented by that in the northern Bering Sea. If the effects of entrainment are also added, the injection rate of water more saline than 33.5 into the Arctic Ocean pycnocline could from this one area alone easily exceed $1 \times 10^6 \text{ m}^3 \text{ s}^{-1}$.

This more detailed examination of the Chukchi and northern Bering seas suggests that in the other shelf seas there may also be regions of ice divergence and enhanced salinization, and that such regions may prove to be large in the aggregate. It is therefore likely that the calculations based on our salinization model (Figs 8 to 12) underestimate the areas capable of contributing water to the Arctic Ocean halocline. Certainly, to assess more fully the role of winter production of dense water on the shelves, the kinematics of the ice must also be considered. A step in this direction has recently been taken by HIBLER (1979).

Fig. 15. Satellite photograph (visible band) of the Chukchi and northern Bering seas, 6 March, 1978. The darker areas of the sea (e.g., south of St. Lawrence Island) represent essentially ice-free conditions.



Fig. 15. Caption on p. 540.

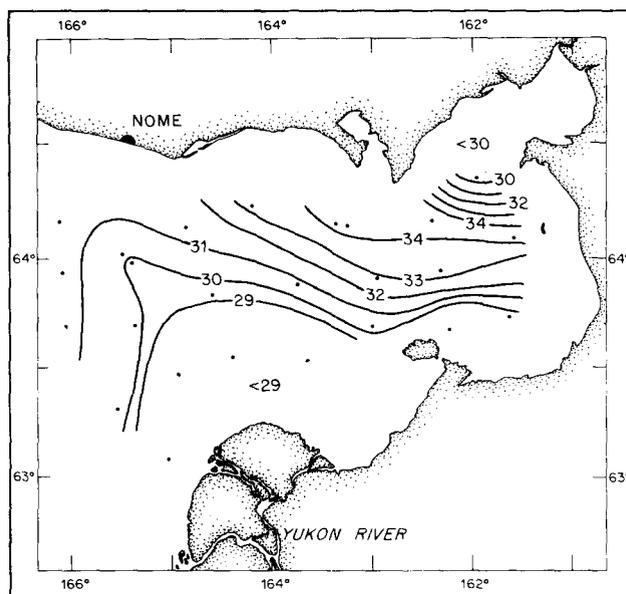


Fig. 16. Near-bottom salinity distribution in Norton Sound, July, 1977.

CONCLUSION

We have argued that the cold, upper halocline of the Arctic Ocean, which underlies a surface mixed layer, is maintained by large-scale lateral advection from certain of the broad continental shelves bordering the Polar Basin. In particular, the area from Spitsbergen to Severnaya Zemlya appears to supply the Eurasian Basin, while the Chukchi and northern Bering seas, possibly supplemented by the region north of the Canadian Arctic Archipelago, feeds the Canadian Basin. The cold and saline shelf water in these areas is produced during freezing and the salinization of the water column is especially pronounced in selected areas of ice divergence, primarily wind-driven. The annual rate at which the cold and saline shelf water feeds into the Polar Basin appears to be in the neighborhood of $2.5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$.

FOSTER (1978) compared the Arctic and Southern oceans and noted a qualitative resemblance in the shape of the T-S correlation diagrams. In both oceans there is a marked change in the gradient of the correlation above the warm-water stratum. Here we have discussed this feature in the Arctic Ocean as caused by the offshore flow of shelf water, whereas MOSBY (1934) and others argued that the analogous feature in the Southern Ocean represents a remnant of local haline convection during winter, the so-called Winter Water. We see no conflict between these two explanations, for there is a difference in the vertical salinity distributions in the two cases. In the Southern Ocean the salinity gradient *increases* below the temperature-minimum layer, whereas in the Arctic Ocean the gradient *decreases*. This is precisely what would be expected for, on the one hand, the remnant of a winter surface mixed layer, the convective depth of which is limited by an underlying halocline, and on the other hand a pre-existing halocline that is thickened by lateral injection of saline water.

As the cold, dense waters flow across the arctic shelves, the rotational effect will deflect

them until geostrophic equilibrium is reached (compare with FOSTER and CARMACK, 1976; KILLWORTH, 1977). Therefore, the actual feeding of the shelf water into the interior ocean probably depends upon bottom Ekman transport or channeling by coastlines and submarine bathymetry. Once the cold, saline water crosses the shelf break, it must eventually sink to the depth where its density matches that of the ambient oceanic waters, at which point it will detach from the bottom and spread into the interior ocean. This is essentially the process described by CARMACK and KILLWORTH (1978) for abyssal waters in the Antarctic. Certainly it appears that the means by which the Arctic Ocean halocline is maintained are very similar to those responsible for the formation of Antarctic Bottom Water (see also GILL, 1973). The chief difference apparently is that the arctic halocline water is insufficiently dense to reach the bottom. Whether or not *any* water from the arctic shelves reaches the abyss is an important question that remains to be answered (see AAGAARD, 1981, for possible evidence).

The significance of the shelf processes that we have described is not that they are responsible for the occurrence of an arctic halocline, for that is due to river discharge and ice melt, but rather that they yield a thicker and colder halocline than would otherwise occur. The thick stratified layer effectively shields the warm Atlantic water from mechanical stirring and winter convection. Conversely, the halocline insulates the surface ice cover from the sub-surface sensible heat by limiting the upward heat flux. (Note particularly that the absence of a temperature gradient in the upper halocline precludes a diffusive heat flux.) Effectively then, the cold halocline, which is fed by shelf water, is also a major heat sink for the Atlantic water (compare with MCPHEE, 1980) and the steady-state heat balance can conceivably be maintained by the equal and opposite effects of lateral advection and vertical diffusion. In such a construction there is no need, or at least a lesser need, for hypothesizing a large local heat loss through the surface as has been suggested, for example, by AAGAARD and GREISMAN (1975); this has consequences for the thermodynamic modeling of the arctic sea ice (see MAYKUT and UNTERSTEINER, 1971). From another viewpoint, if the salinity of shelf water were to be increased, for example, by massive river diversion projects (see AAGAARD and COACHMAN, 1975; GRIBBIN, 1979), water now maintaining the halocline might become sufficiently dense to sink into the abyss, thus thinning the pycnocline and making the Atlantic layer more susceptible to direct surface cooling. Undoubtedly this would have consequences for the ice distribution.

We are led by all these considerations to the conclusion that shelf processes are of essential importance in understanding the oceanography of the Arctic.

Acknowledgements—We are greatly indebted to C. H. DARNALL, S. HARDING, J. H. SWIFT, and R. B. TRIPP for their efforts in the field. Part of this work was supported by the Bureau of Land Management through inter-agency agreement with the National Oceanic and Atmospheric Administration, under which a multi-year program responding to needs of petroleum development of the Alaskan continental shelf is managed by the Outer Continental Shelf Environmental Assessment Program (OCSEAP) office. Other financial support came from the Office of Naval Research under contract N00014-75-C-0893.

REFERENCES

- AAGAARD K. (1977) STD measurements in possible dispersal regions of the Beaufort Sea. Annual Report, Contract 03-5-022-67, TO1. Department of Oceanography, University of Washington, 34 pp. (Unpublished document.)
- AAGAARD K. (1978) STD measurements in possible dispersal regions of the Beaufort Sea. Annual Report, Contract 03-5-022-67, TO1. Department of Oceanography, University of Washington, Ref. A78-10, 14 pp. (Unpublished document.)

- AAGAARD K. (1979) Current measurements in possible dispersal regions of the Beaufort Sea. Annual Report, Contract 03-5-022-67, TO3. Department of Oceanography, University of Washington, Ref. A79-5, 23 pp. (Unpublished document.)
- AAGAARD K. (1981) On the deep circulation in the Arctic Ocean. *Deep-Sea Research*, **28**, 251–268.
- AAGAARD K. and L. K. COACHMAN (1975) Toward an ice-free Arctic Ocean. *EOS*, **56**, 484–486.
- AAGAARD K. and P. GREISMAN (1975) Toward new mass and heat budgets for the Arctic Ocean. *Journal of Geophysical Research*, **80**, 3821–3827.
- CARLETON A. M. (1980) Polynya development in the Cape Thompson–Point Hope Region, Alaska. *Arctic and Alpine Research*, **12**, 205–214.
- CARMACK E. C. and P. D. KILLWORTH (1978) Formation and interleaving of abyssal water masses off Wilkes Land, Antarctica. *Deep-Sea Research*, **25**, 357–369.
- COACHMAN L. K. and K. AAGAARD (1974) Physical oceanography of arctic and subarctic seas. In: *Marine geology and oceanography of the Arctic Seas*, Springer-Verlag, New York, pp. 1–72.
- COACHMAN L. K., K. AAGAARD and R. B. TRIPP (1975) *Bering Strait: the regional physical oceanography*, University of Washington Press, Seattle, 172 pp.
- COACHMAN L. K. and C. A. BARNES (1962) Surface water in the Eurasian Basin of the Arctic Ocean. *Arctic*, **15**, 251–277.
- FOSTER T. D. (1978) Polar oceans: similarities and differences in their physical oceanography. In: *Polar research in the present, and the future*, M. A. McWHINNIE, editor, AAAS Selected Symposium 7. Westview Press, pp. 117–140.
- FOSTER T. D. and E. C. CARMACK (1976) Frontal zone mixing and Antarctic Bottom Water formation in the southern Weddell Sea. *Deep-Sea Research*, **23**, 301–317.
- GILL A. E. (1973) Circulation and bottom water production in the Weddell Sea. *Deep-Sea Research*, **20**, 111–140.
- GRIBBIN J. (1979) Climatic impact of Soviet river diversions. *New Scientist*, **84**, 762–765.
- HIBLER W. D. (1979) A dynamic thermodynamic sea ice model. *Journal of Physical Oceanography*, **9**, 815–846.
- HUANG J. C. (1972) The thermal bar. *Geophysical Fluid Dynamics*, **3**, 1–25.
- KILLWORTH P. D. (1977) Mixing on the Weddell Sea continental slope. *Deep-Sea Research*, **24**, 427–448.
- KINNEY P., M. E. ARHELGER and D. C. BURRELL (1970) Chemical characteristics of water masses in the Amerasia Basin of the Arctic Ocean. *Journal of Geophysical Research*, **75**, 4097–4104.
- MALMGREN F. (1927) On the properties of sea ice. *The Norwegian North Polar Expedition with the "Maud"*, 1918–1925. *Scientific Results*, **1**(5), Bergen, 85 pp.
- MAYKUT G. and N. UNTERSTEINER (1971) Some results from a time-dependent, thermodynamic model of sea ice. *Journal of Geophysical Research*, **76**, 1550–1575.
- MCPHEE M. G. (1980) Oceanic heat flux in the Arctic: a peculiar thermohaline regime. *Ocean Modelling*, **31**, 1–4.
- MOSBY H. (1934) The waters of the Atlantic Antarctic Ocean. *The Norwegian Antarctic Expeditions, 1927–1928. Scientific Results*, **1**(11), 131 pp.
- MOUNTAIN D. G., L. K. COACHMAN and K. AAGAARD (1976) On the flow through Barrow Canyon. *Journal of Physical Oceanography*, **6**, 461–470.
- NANSEN F. (1902) Oceanography of the North Polar Basin. *The Norwegian North Polar Expedition, 1893–96. Scientific Results*, **3**(9), 427 pp.
- NEUMANN G. and W. J. PIERSON, JR. (1966) *Principles of physical oceanography*. Prentice-Hall, New Jersey, 545 pp.
- SHIRSHOV P. P. (1944) Scientific results of the drift of station North Pole. *Akademiya Nauk, SSSR, Obshchee Sobranie*, February 1944, pp. 110–140 (Translation).
- SVERDRUP H. U. (1929) The waters on the north Siberian shelf. *The Norwegian North Polar Expedition with the "Maud"*, 1918–1925. *Scientific Results*, **4**(2), Bergen.
- THORNDIKE A. S., D. A. ROTHROCK, G. A. MAYKUT and R. COLONY (1975) The thickness distribution of sea ice. *Journal of Geophysical Research*, **80**, 4501–4513.
- TIMOFEYEV V. T. (1951) Resultaty glubokovodnikh gidrologicheskikh nabludenii. *Trudy dreifzuyushchi ekspeditsii Glavsemorputi na lp "G. Sedov"*, 1937–1940, Vol. 4, Izdatel'stvo Glavsemorputi, Moscow–Leningrad.
- TIMOFEYEV V. T. (1960) *Vodnye Massy Arkticheskogo Basseina*. Gidrometeoizdat, Leningrad, 190 pp.
- TRESHNIKOV A. F. (1959) Poverkhnostniye vodi v arkticheskome basseine. *Problemy Arktiki*, **7**, 5–14.
- TRESHNIKOV A. F. and G. I. BARANOV (1972) *Struktura tsirkulyatsii vod Arkticheskogo Basseina*. Gidrometeoizdat, Leningrad, 158 pp.
- TRIPP R. B. and K. KUSUNOKI (1967) *Physical, chemical, and current data from ARLIS II: eastern Arctic Ocean, Greenland Sea, and Denmark Strait areas, February 1964–May 1965*, Vol. 1, Technical Report No. 185, Department of Oceanography, University of Washington, 341 pp. (Unpublished document.)
- WISEMAN W. J. (1977) Hypersaline bottom water and permafrost degradation. *EOS*, **58**, 1152 (abstract).