

Geochemical Profiles in the Central Arctic Ocean: Their Relation to Freezing and Shallow Circulation

R. M. MOORE AND MALCOLM G. LOWINGS

Department of Oceanography, Dalhousie University, Halifax, Nova Scotia, Canada, B3H 4J1

F. C. TAN

Atlantic Oceanographic Laboratory, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada, B2Y 4A2

Temperature, salinity, nutrient, tritium, and oxygen isotope data collected during the Lomonosov Ridge Experiment along a drift track between the Makarov and Fram Basins over the Lomonosov Ridge are presented. The relationship of these quantities to the processes that maintain the halocline, in particular to the production of more saline waters by addition of brines formed during the freezing of seawater, is described. The results support the idea that the wide continental shelves of the Arctic Ocean play an important role in maintaining the halocline.

INTRODUCTION

Questions of particular interest pertaining to the Arctic Ocean relate to its important role as a region of heat exchange with the atmosphere, to the effects that ice cover has on mixing of the surface layer and on heat and gas exchange, to the unusual light cycle and its effect on productivity, and to the chemical and physical effects of freezing sea water. A comparison of this ocean, with its low productivity (at least during winter), with other oceanic regions should help to elucidate some of the effects of biological activity on marine chemistry, for example, trace metal cycling and speciation.

Aagaard et al. [1981] discussed the way in which the cold waters of the halocline which lie immediately below the surface mixed layer and extend to the top of the warm Atlantic layer could be produced on the wide continental margins surrounding the Arctic Ocean. The processes involve freezing with release of highly saline brines having a temperature close to the freezing point of surface sea water. This water flows over the shelves and into the central basin, sinking then to the appropriate density surface. The distributions of various chemical parameters will be affected in predictable ways by these processes.

As a part of the Lomonosov Ridge Experiment (LOREX), which ran from late March until the end of May 1979, a program was undertaken aimed at describing more accurately the hydrography of the central Arctic Ocean and studying the geochemistry of selected trace elements in this unusual environment. To this end, water samples were collected from a drifting ice station (LOREX satellite camp 1) for the measurement of salinity, silicate, phosphate, $\delta^{18}\text{O}$, trace metals [Moore, 1981, 1983], carbon-14, tritium, helium-3, caesium-137, caesium-134, strontium-90, plutonium, americium (H. D. Livingston et al., manuscript in preparation, 1983), alkalinity, and pH. Ice cores were collected to investigate the possibility of deducing the extent of temporal changes in the tritium content of the surface waters.

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During the period of sampling, April 13 to May 21 (days 103-141), the camp drifted from 88°40'N 139°50'W to 89°9'N 97°7'W. The drift track (Figure 1) extends from the Makarov Basin, a part of the Amerasian Basin with a water depth of ca. 4000 m, across the Lomonosov Ridge, where the depth decreased to about 1000 m, into the Fram Basin and depths of ca. 2500 m.

METHODS

Temperatures were measured using reversing thermometers on 1.7-l Niskin and 5-l Go-Flo bottles, and salinities were determined by using a portable Bissett-Berman inductively coupled salinometer.

Reactive silicate was determined spectrophotometrically by the method described by *Strickland and Parsons* [1972], involving a metol-reduced silicomolybdate complex. The determinations were normally carried out within about 24 hours of sample collection. Reactive phosphate was determined by the method of *Strickland and Parsons*, the samples being frozen immediately after collection and determined for phosphate within a few days. A Varian model 634 spectrophotometer was used for both of these analyses.

Measurements of $\delta^{18}\text{O}$ were made according to the methods given by *Tan and Strain* [1980], and tritium measurements were made by H. G. Ostlund [*Ostlund*, 1982].

TEMPERATURE AND SALINITY

The first profile was sampled during the period April 16-29 (days 106-119) and was marked initially by very slow movement of the ice. The surface layer to a depth of about 50 m (Figure 2) is of low salinity, ca. 30 ‰, but fluctuations between 30.059 and 30.409 ‰ are apparent: because of the very few samples involved, no explanation will be proposed for this anomaly.

The second profile (Figure 2) comprises samples collected in the period May 3-15 (days 123-135). The near-surface salinities recorded during a period of more rapid ice movement (day 123) show an almost homogeneous layer to about 40 m. There is a slight decline in salinity from 30.152 ‰ at 5 m to 30.137 ‰ at 25 m consistent with an addition of salt at the surface: the salinity increases a small amount to 30.175 ‰ at 40 m. At about 40 m a steep rise in salinity begins (as much as 0.13 ‰ per m at 40 m) reaching 34 ‰ by 130 m. The

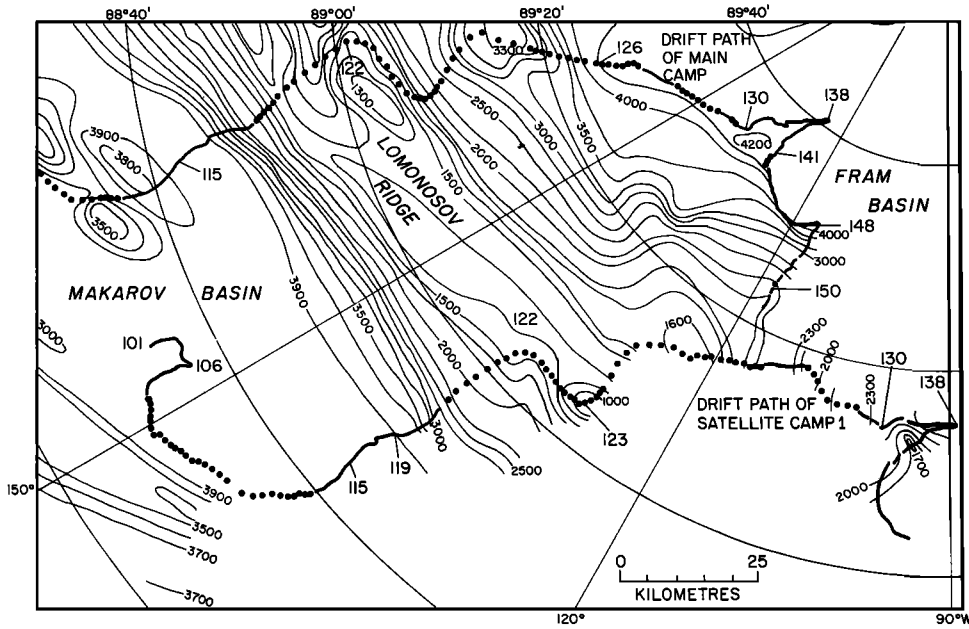


Fig. 1. LOREX satellite camp 1 drift path (Julian days indicated).

surface salinities reported here are at the lower limit of the range (30–32 ‰) reported by *Coachman and Barnes* [1962] and *Treshnikov et al.* [1977] for the central Arctic Ocean.

Between the bottom of the mixed layer and about 200 m the two profiles have a similar shape, but the salinities of the more westerly profile are displaced to greater depths by 10–15 m. At the core of the Atlantic water (ca. 450 m) as defined by the temperature maximum (Figures 3 and 4), the salinity was 34.89 ‰. Although there was generally good agreement between the salinities of different samples from similar depths in the deep water [*Aagaard*, 1981], there is some

scatter at around 1500 m with salinities in the range 34.91–34.94. The salinities in this region have been dealt with in some detail by *Aagaard* [1981]. The number of salinity measurements below 2000 m is rather small, but they indicate that at least between 2000 and 2500 m the Makarov Basin salinities are slightly greater (by 0.01 ‰) than those of the Fram Basin at the same depth.

The T-S diagram (Figure 3) shows that the temperature remains constant and near freezing point to a depth of about 100 m, it then increases slightly over the depth range 100 to 160 m. At this point the diagram shows a marked inflection point and the temperature begins to rise sharply to a prominent maximum which is the characteristic of water originating in the Atlantic. This inflection point corresponds to that reported by *Kinney et al.* [1970] at T3 (position ca. 80°–83°N, 160°W) at a depth of 230 m but at the same temperature and density, $\sigma_T = 27.6$. *Aagaard et al.* [1981] have pointed out that the waters lying between the mixed layer and the

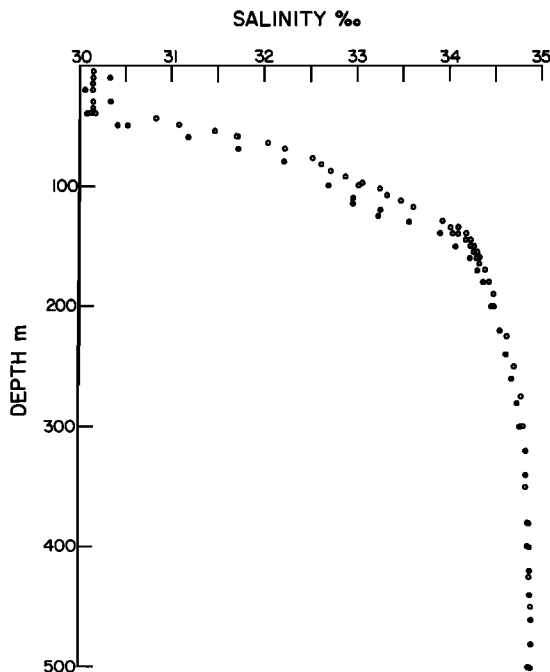


Fig. 2. Salinity versus depth at LOREX station. Solid circles, profile 1; open circles, profile 2.

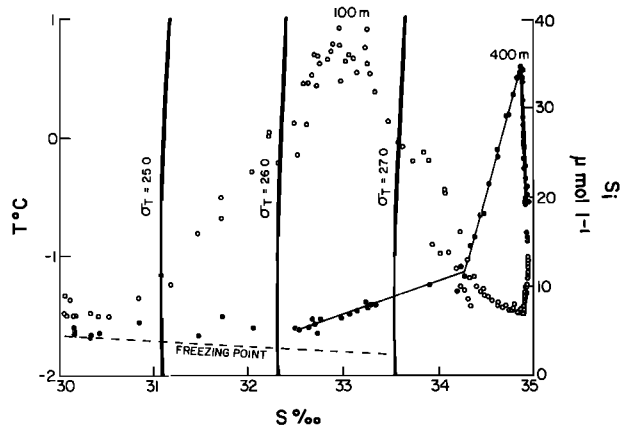


Fig. 3. Temperature-salinity relationship (solid circles) and silicate-salinity relationship (open circles) at the LOREX site; data from both profiles are included. Broken line indicates freezing point at atmospheric pressure.

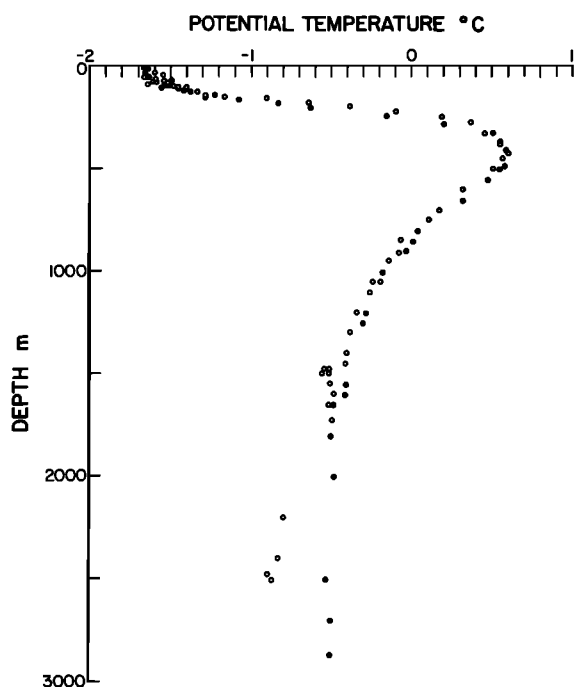


Fig. 4. Potential temperature, surface to 3000 m. Solid circles, profile 1; open circles, profile 2.

temperature maximum cannot be produced simply by mixing surface water (0–50 m) and the relatively warm Atlantic water, and they suggest that these waters are produced over the shelves by a process involving freezing and salt addition. In the discussion of the silicate, tritium, and stable oxygen isotope ratios, it will be shown that their distributions are in accordance with this interpretation.

Coachman and Barnes [1962] give an alternative explanation of how cold saline water of the type described above may be formed by a process involving the submarine canyons of the continental slope which provide access for Atlantic water to the shelf regions where mixing with low salinity water and cooling occurs.

It is worth noting that the temperature maximum that is found in much of the Canada Basin in the halocline [*Coachman and Barnes*, 1961] is scarcely discernible at this region (Figures 3 and 5), at least during late winter.

Below 2000 m the two potential temperature profiles diverge (Figure 4), the higher temperatures occurring in the Makarov Basin. Data are available to 2900 m and 2500 m in the Makarov and Fram Basins, respectively; the temperature difference between the two basins becomes discernible at 2000 m and is marked at 2500 m.

SILICATE AND PHOSPHATE

While the T-S diagram (Figure 3) gives no indication of the presence of a well-defined water mass centered at about 100 m, the silicate profiles (Figures 6 and 7) show a well-defined maximum at this depth with silicon concentrations of almost $40 \mu\text{mol l}^{-1}$. Beneath the peak the concentration falls rapidly, until at about 250 m it is near to the surface water value of $7 \mu\text{mol l}^{-1}$. Below 800 m (Figure 7), concentrations increase again with depth reaching a maximum value of $13 \mu\text{mol l}^{-1}$ in the Makarov Basin at a depth of ca. 2300 m, then remaining almost constant to 3000 m (bottom depth 3380 m). At around 1800 m a marked divergence appears between the

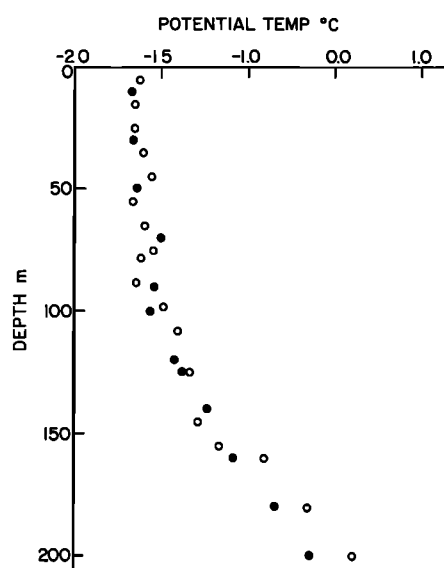


Fig. 5. Potential temperature, surface to 200 m. Solid circles, profile 1; open circles, profile 2.

two profiles, the concentration of silicate decreasing with further increase in depth in the Fram Basin to a minimum at 2200 m followed by a small increase toward the bottom, which at this position was at a depth of 2600 m. The maximum in silicate at about 1700 m in the profile in the Fram Basin is apparently supplied by water in the Makarov Basin.

The silicate data, when plotted all together as in Figure 3, show considerable scatter; this is due to the presence of steep lateral gradients. If only those samples that were collected simultaneously are plotted together, well-defined relationships become apparent. In Figure 8, silicate values are plotted in this way against salinity for the depth range from just below the surface mixed layer to beneath the nutrient maximum.

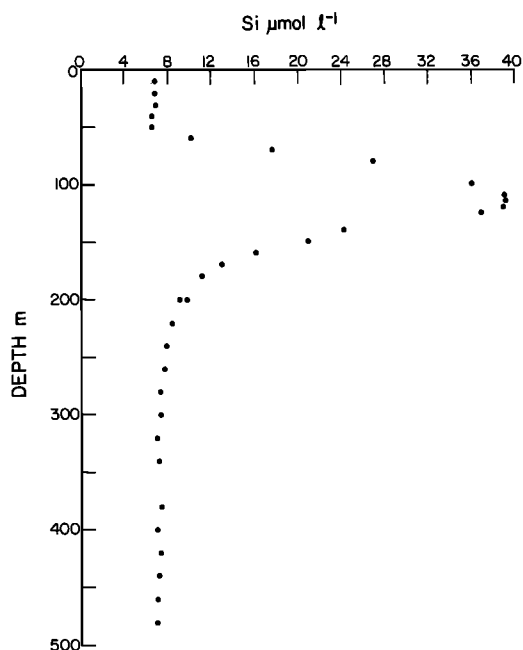


Fig. 6. Silicate against depth, surface to 500 m, profile 2.

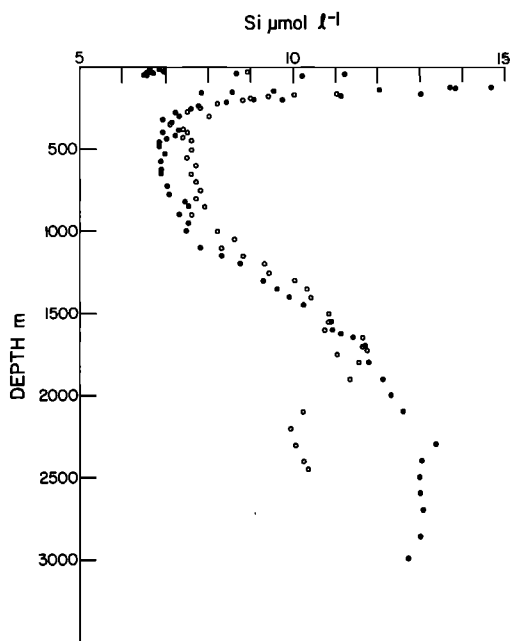


Fig. 7. Silicate against depth to 3000 m. Solid circles, profile 1; open circles, profile 2.

Considering first the waters lying between the surface mixed layer and the nutrient maximum, it is seen that generally the silicate and salinity values are appropriate to mixtures of waters from the nutrient maximum and water from just beneath the mixed layer rather than with the mixed layer itself. The Si-S characteristics of the mixed layer water (Figure 3) are appropriate to a mixture of the underlying layer with a fresh component, comprising meltwater and runoff, with the lower salinity of the surface layer demanding a freshwater component equivalent to only ca. 1% of its volume. Consequently, while the silicate concentration of runoff is unknown and could lie within a wide range [Codispoti and Lowman, 1973], the actual value will have little effect on the silicate concentration of the surface water. The water that lies immediately below the mixed layer may be expected to have a component of brine derived from freezing sea water. Figure 8 shows that the linear Si-S relationships above the nutrient maximum can have quite different slopes. The steeper slope can be accounted for by the introduction of water having similar Si concentration but higher salinity than that at 50 m, i.e., having a greater brine component than the ambient waters of similar silicate concentration. It will be seen that the tritium level of this water is also consistent with it being formed with a large brine component.

The water of the nutrient maximum is well established in the literature as coming from the Chukchi Sea region [Kinney *et al.*, 1970]. The elevated nutrient concentrations are then simply a reflection of the source of these waters in the Bering Sea; it is suggested here, however, that the prominence of this water type is the result primarily of the presence of the nutrient marker and that waters of similar salinity and temperature, hence density, are also produced in other continental shelf locations from where they flow to lie above, below, and perhaps to interleave the nutrient maximum water. It is worth noting that any waters that have been enriched in salinity by freezing in the marginal seas and flowed over the shallow shelf sediments to enter the central

basin will have tended to accumulate nutrients that diffuse from the sediment surface. Such a process is likely to be enhanced in the Arctic Ocean on account of the unusually extensive continental shelves; it would be further enhanced by a strong density gradient near the bottom of the shelf seas, as this could confine the nutrients of sedimentary origin to a relatively thin layer of water and produce a correspondingly higher anomaly. Thus it may be the case that the nutrient-rich waters of the halocline have a wider origin than is immediately apparent. This would help to explain the observation that the temperature maximum which is normally attributed to water that has entered the Arctic Ocean from the Bering Strait during summer decays away more rapidly than the nutrient signal of the underlying water, the former being neutralized by cold water introduced at the same density, while the latter is weakened only to the extent that the waters added from other sources have lower nutrient levels.

Looking now at the waters lying beneath the Si maximum (Figure 3), it is seen that they mix with some water that is low in silicate and is considerably cooler than the Atlantic water. From Figure 9 it is seen that two Si profiles show a nonlinear Si-S relationship: the silicate maximum is one end-member, and the Atlantic water with its minimum silicate values constitutes a second component. There must be another component that produces this curvature, and one series of samples (Figure 9, crosses) clearly indicates a body of water having low Si concentrations but at a significantly lower salinity (34.3 ‰) than the Atlantic water (34.9 ‰) and also at much lower temperature. It is this water that marks the inflection point on the T-S diagram, the temperature rising steeply below this point (Figure 10). The similarity between this water and the Atlantic water in silicate concentration suggests that it may be modified Atlantic water. *Coachman and Barnes* [1962] have suggested how Atlantic water may be modified in temperature and salinity by a process involving submarine canyons on the continental shelves. *Aagaard et al.* [1981] have further commented on this process.

Phosphate data, available only for the Makarov Basin, and not detailed here, form a profile essentially like that of silica, though small variations in the water column are less readily resolved on account of the lower precision of the analysis.

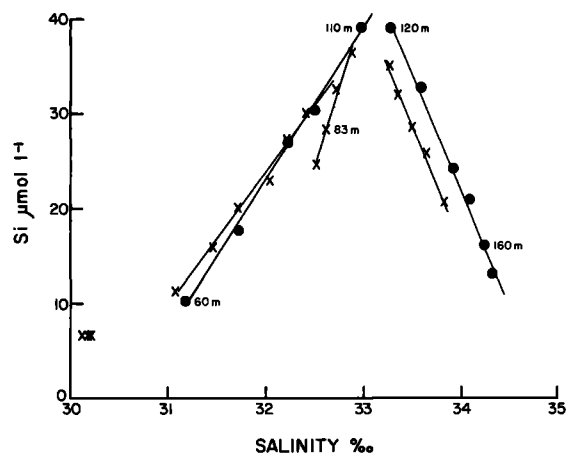


Fig. 8. Silicate versus salinity, 50–150 m. Solid circles, samples collected days 106–108; crosses, days 123–125.

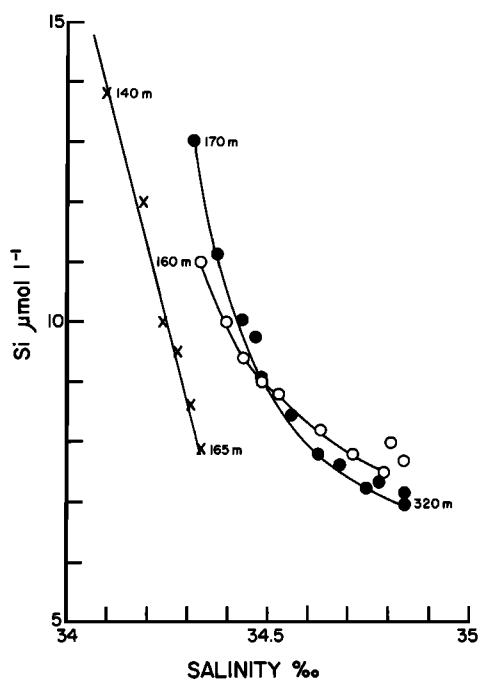


Fig. 9. Silicate versus salinity, 140–325 m. Solid circles, samples collected days 108–109; open circles, day 99; crosses, day 95.

An appendix giving raw data is available on microfiche.¹ The phosphate and silicate maxima are coincident in depth, as expected for an advective feature.

TRITIUM

Tritium samples were collected throughout the duration of the project; the data described here are primarily from the upper 200 m of the water column. For treatment of the deeper samples reference should be made to Östlund [1982].

The data are shown in Figure 11. Tritium values for the depth range 200–3000 m are essentially consistent with the presence of a freshwater component having a tritium concentration of 675 TU [Östlund, 1982]. Figure 11 shows that between 50 and ca. 200 m tritium varies less steeply with salinity; extrapolation of a straight line through these points would yield a freshwater component containing ca. 150 TU. There are a number of samples containing somewhat higher tritium, and these will be dealt with separately. The mixed layer contains a markedly higher tritium concentration; this will be dealt with first.

In the same way that the silicate data in the surface mixed layer were interpreted as the result of an addition of a fresh component to water of the subsurface type, so do the tritium values fit this interpretation. But in this case it must be assumed that it is the meteoric component of the freshwater that accounts for the bulk of the tritium enhancement. The meltwater component would be expected to have a tritium level similar to that of the surface water from which it is ultimately derived, i.e., 30–35 TU [Weston, 1955]. Ice cores from the LOREX site confirm this estimate; though a few

¹ Appendix table is available with entire article on microfiche. Order from the American Geophysical Union, 2000 Florida Avenue, N.W., Washington, D. C. 20009. Document C82-001; \$2.50. Payment must accompany order.

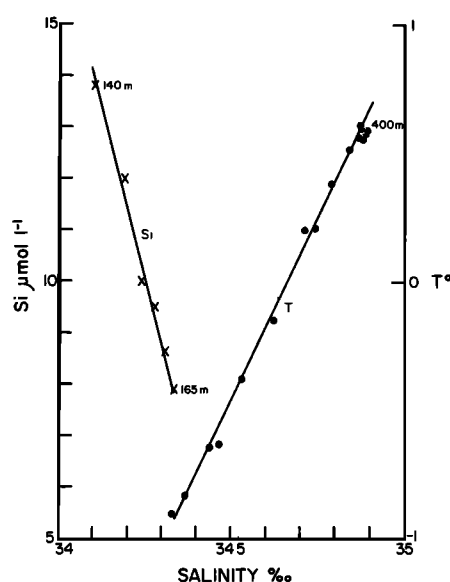


Fig. 10. Silicate versus salinity, 140–165 m, samples collected day 95 and temperature versus salinity for all samples 140–480 m.

sections showed higher levels [Östlund, 1982], these are likely to contain a fraction of meteoric water which would be present on the surface as snow but could also be present within the core as a result either of freezing of meteoric water on the underside of the ice or of percolation of rain or melted snow through the ice during summer.

If the 50 m sample, $S = 31.080$ ‰, is taken as representative of the water immediately beneath the mixed layer and a tritium value is then calculated for the freshwater that must be added to give the mixed layer salinity, we get a value of ca. 450 TU. Then taking a figure of 675 TU for the meteoric component [Östlund, 1982] and 35 TU for meltwater, an estimate can be made of the proportions of each type needed to give a mixture of 450 TU. This calculation indicates a mixture of 35% meltwater and 65% meteoric water. This is in tolerable agreement with the ratio that would be calculated for the entire Arctic Ocean by using the value for precipitation and runoff of 0.45 m yr^{-1} [SCOR Working Group 58, 1979] and of 0.6 m yr^{-1} for ice melting [Thorndike et al., 1975].

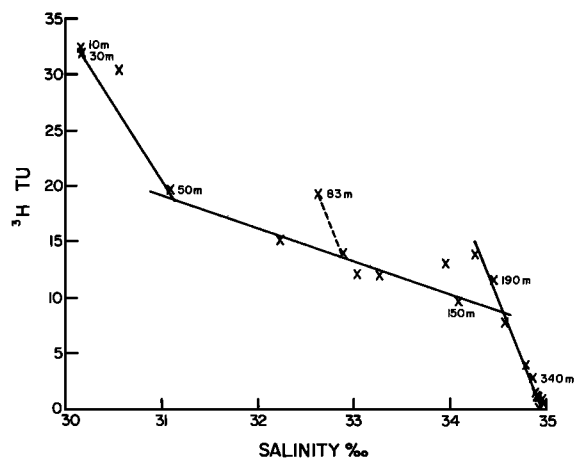


Fig. 11. ³H versus salinity.

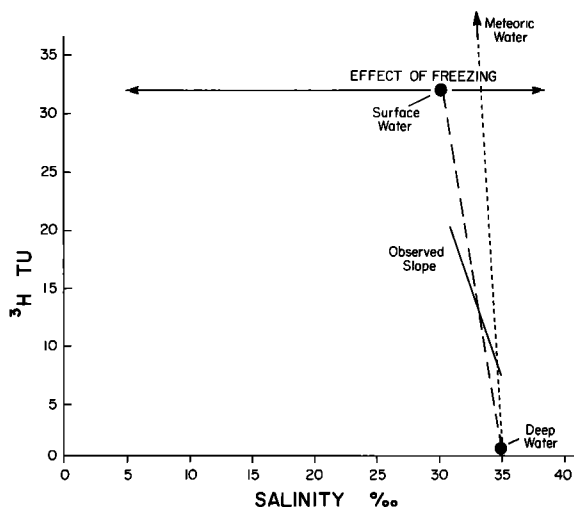


Fig. 12. Illustration of the way in which freezing and melting of surface water yields two components that can reduce the slope of the ^3H -S relation below that obtainable by mixing the three components deep, surface, and meteoric water.

A slightly more refined calculation takes into account the fact that the 50-m sample is slightly more saline than the water type that is contributing to both the overlying mixed layer and the underlying waters down to the nutrient maximum; this is demanded by the silica concentration of the mixed layer and of freshwater. The result is similar: the tritium of the freshwater addition becomes 480 TU, which is equivalent to 70% meteoric water and 30% meltwater.

It is now necessary to account for the rather slow decrease in tritium with salinity below the mixed layer to ca. 200 m. This is essentially due to the fact that these waters contain a significant component of laterally advected water, formed on the continental shelf from shallow, tritium-rich water to which brine has been added.

The slope of the ^3H -S line is then determined by the proportion of meteoric water and meltwater added with modification due to the addition of brine having a tritium content similar to that of surface water (ca. 35 TU). The plot of tritium against salinity (Figure 11) suggests that this process is significant to a depth of about 160–200 m, so it matches the depth range over which the T-S plot indicates advection of cool waters into the Arctic Basin.

The process described here is illustrated in Figure 12. The gradient of the ^3H -S line resulting from mixing surface and deep water is shown; also indicated is the steeper slope that would result from mixing meteoric water with deep water. It is observed, however, that an intermediate depth range alone exhibits a shallower gradient than these components can accommodate. Freezing and melting of the surface water produces two new components that have the appropriate properties to yield the observed gradient, provided that they are mixed in at different depths as their densities would demand.

The few samples that have anomalously high tritium levels represent traces of advected layers which have not yet completely mixed with the ambient water of the same density, so they retain a signature of elevated tritium and also a remnant of their original nutrient concentration. In the specific case of the sample from 83 m, this appears (Figure 8) as a low silicate value compared with the background of

nutrient rich water having its origin in the Bering Sea. Also, between 150 and 200 m there are a few samples that have relatively high tritium levels. These are apparently associated with the water that advects at the density surface corresponding to the inflection point on the T-S plot. Water at this level (corresponding to about 160 m) has a tritium content of ca. 13 TU; this represents only a small decrease from the level at around 70 m even though the salinity has increased by ca. 2 ‰.

Below ca. 200 m the ^3H -S plot has a slope appropriate to the addition of meteoric water only, the ^3H -S relationship in this region being established outside the Arctic Ocean. Meltwater is not apparent at this location below a depth of about 200 m.

STABLE OXYGEN ISOTOPES

Since the processes that have been invoked to account for the distribution of T, S, Si, and tritium have an influence on the stable oxygen isotope composition of sea waters, it should be possible to relate the foregoing arguments to the $^{18}\text{O}/^{16}\text{O}$ ratios in the water column.

Measurements of $\delta^{18}\text{O}$ in Arctic Ocean waters have been reported by *Vetshteyn et al.* [1974] and *van Donk and Mathieu* [1969]. Owing to either the data being too sparse, particularly in the upper waters (0–400 m) or the vertical profiles being made up of samples from widely different sampling stations, these authors are unable to examine the processes of freezing and meteoric water addition that would affect the $\delta^{18}\text{O}$ distributions. In addition to the significance of $\delta^{18}\text{O}$ data in relation to the processes that have been discussed above, the presentation of detailed $\delta^{18}\text{O}$ data for the upper few hundred meters of the water column is also important in relation to Arctic paleoclimatology and paleoceanography. In order to understand the $\delta^{18}\text{O}$ record preserved by fossil forams, it is essential to have some knowledge of the present Arctic surface water and the relationships between $\delta^{18}\text{O}$ and salinity.

The data are shown in Figure 13. Compared with surface mixed layer water, the water lying immediately beneath (at 60 m) shows a similar $\delta^{18}\text{O}$ but higher salinity; this is in accordance with its formation by brine addition to a surface water type. Below this depth we should expect a relationship between $\delta^{18}\text{O}$ and salinity consistent with formation of the

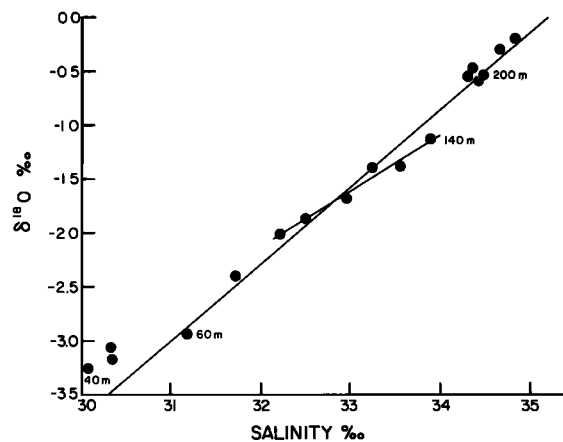


Fig. 13. $\delta^{18}\text{O}$ versus salinity, depth range surface to 320 m. Equation of line through all samples, $\delta^{18}\text{O} = -25.2 + 0.71S$. Equation of lines through samples in depth range 80–140 m, $\delta^{18}\text{O} = -18.6 + 0.52S$.

waters from a surface type with a component of brine and a component of the underlying water. It is possible to estimate the form of the $\delta^{18}\text{O}$ -salinity relationships that would be observed in the extreme cases of (1) salinity modification by brine addition alone and (2) by mixing of surface water and water from beneath the halocline. Case (1) would have a gradient $d(\delta^{18}\text{O})/dS$ of near zero because freezing produces only a small fractionation of oxygen isotopes. Case (2) would have a gradient dependent on the proportion of meteoric and meltwater being added at the surface. This proportion should be similar to the ratio that was estimated to account for the difference in tritium between surface and subsurface waters. Taking a mixture of 70% meteoric water ($\delta^{18}\text{O} = -30$) and 30% meltwater ($\delta^{18}\text{O} = -3$) would give a $\delta^{18}\text{O}$ for the mixture of -22 . If water of this type were allowed to mix with a saline end-member having a salinity of 34.5 ‰ (the value chosen is not critical, but is selected as a likely lower limit for freshwater penetration), the equation relating $\delta^{18}\text{O}$ and salinity can be predicted as

$$\delta^{18}\text{O} = -22 + 0.634S \text{ ‰}$$

The choice of $\delta^{18}\text{O}$ for the meteoric component has a critical effect on the value of the intercept at $S = 0$. The processes invoked to explain the distribution of geochemical parameters should yield a $\delta^{18}\text{O}$ -salinity relationship between the two limits described by cases (1) and (2), with curvature due to some displacement of points toward higher salinity in those places where addition of brine is occurring.

The data (Figure 13) show some sign of displacement of points to higher salinity around 60 m and 130 m, otherwise they are indicative of dilution of the deeper waters with a mixture of meltwater and meteoric water in amounts similar to those used in the above calculation.

Although the $\delta^{18}\text{O}$ versus salinity plot in Figure 13 does not distinctly illustrate regions of brine addition, it does, by giving the proportion of meltwater present, indicate how much brine has been distributed through the water column. The way in which it becomes distributed is yet to be detailed. Here it is suggested that there is some localization of brine introduction immediately beneath the mixed layer and also at ca. 130–140 m. The question of whether any of the brine is transported to deep water is beyond the scope of this study, but it may be noted that a small enhancement of deep water salinities could much reduce the significance of brine addition in the relatively thin near-surface layers. Aagaard [1981] has discussed the question of salt addition to deep Arctic Ocean waters based on data from the LOREX site.

The difficulty here is trying to identify small variations in the $\delta^{18}\text{O}$ -salinity gradient from a small number of data points having a precision that is low relative to the range of variability existing in the water column.

CONCLUSIONS

Measurements of silicate and tritium in the upper 200 m of the central Arctic Ocean make a valuable addition to temperature and salinity data; they exhibit a complex pattern but one that is consistent with the view that the halocline is supplied by horizontal advection with waters produced on the continental shelves from shallow waters by salinity enhancement with brine. A degree of complexity stems from the lateral inhomogeneity of the water column even at a great distance from the margins of the basin.

In addition to the well-characterized water mass having an

origin in the Chukchi Sea region [Kinney *et al.*, 1970], another deeper contributor to the halocline can be recognized from these data; this water is characterized by low silicate concentrations and is sufficiently similar to the Atlantic layer to suggest that it is derived from those waters. A plausible mechanism for the transformation of the Atlantic water was put forward by Coachman and Barnes [1962].

This work demonstrates the value of chemical measurements in distinguishing between different water types particularly when they lie at the same density surface and have little temperature variation. It follows, too, that the chemical data should be able to give some indication of the rate at which horizontal mixing occurs. Silicate measurements are of particular value in the Arctic Ocean owing to both the range of concentrations in the upper waters and the precision of the measurements. On the other hand, stable oxygen isotope measurements appear not to have the dynamic range necessary to reveal in detail the effects of brine addition on the water column.

It is concluded that in view of the marked lateral inhomogeneity of the water column, it is important to collect samples simultaneously as far as is possible. Also, it is apparent that the greatest value can be derived from chemical measurements when they are made on the same water samples, so that direct comparison of parameters is possible rather than the less satisfactory comparison of profiles.

Future work on the chemistry of brines might result in sensitive tests for the extent of their influence on high-latitude water masses. Apart from their intrinsic chemistry, they may be further characterized by virtue of chemical modifications that occur while they lie in contact with the sediment surface, for example, losing oxygen and accumulating components that diffuse from the sediments: nutrients, radon-222, and radium-228.

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