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A Relationship Between Heat Transfer to Sea Ice and Temperature-Salinity Properties of Arctic Ocean Waters

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It is demonstrated that the characteristic temperature-salinity relationship shown by thermocline waters of the Arctic Ocean can be reproduced using a simple model based on transfer of heat directly from these waters to sea ice. From this we deduce that there is direct sensible heat flux from waters of Atlantic origin to the ice cover, such interaction could play a major role in determining the physical properties of Arctic water masses.

INTRODUCTION

A prominent feature found throughout the Arctic Ocean is a layer of relatively warm water of Atlantic origin centered at a depth of a few hundred meters. Various authors have attempted to account for the origin of the cold halocline waters that lie above this layer [*Coachman and Barnes*, 1962; *Aagaard et al.*, 1981; *Melling and Lewis*, 1982]; there is a tendency for such accounts to be developed from the starting point that the halocline waters cannot be a two-component mixture of Atlantic and surface waters because such mixing would yield intermediate temperatures which are not observed. We suggest that this argument is misleading and that the properties of thermocline waters, at least in the western Arctic, may be explicable in terms of interaction between waters having temperatures above the freezing point and ice which is the true surface layer.

EFFECT OF ICE MELTING ON T AND S

Temperature-salinity (T-S) relationships of the form shown in Figure 1 are typical of the western basins of the Arctic Ocean. Considering that most of the Arctic Ocean experiences a long season of ice-cover, we may enquire what would be the result of mixing waters from the region A-B of the T-S diagram with the surface ice. While the temperature of the lower surface of the ice is typically approximately -1° to -2°C, its influence on temperature during mixing is determined almost entirely by the large latent heat of fusion of ice L (measured in joules per kilogram).

The latent heat of sea ice is a function of its salinity and temperature, being determined by the proportion of water present as ice or as entrapped brine. Sea ice has a range of salinities averaging ~6 (6-8 near the ice-water interface) [*Nakawo and Sinha*, 1981]. As we are considering ice at or near the water interface, a value of 7 will be adopted here, but it will be shown that choice of salinity is not critical to our argument. (It should be noted that references to "salinity of ice" actually mean salinity of water derived from melting ice.) The latent heat of sea ice of this salinity and having an appropriate temperature of -1.8°C is 2.67 x 10^5 J kg^{-1} [*Ono*, 1967].

The equation that relates the changes in salinity and temperature resulting from mixing M_i g ice and M_w g seawater, having initial salinities S_i and S_w, temperatures T_i and T_w, and densities ρ_i and ρ_w, is derived as follows. The salinity of the mixture S_m is

$$\frac{(M_i + M_w)}{\rho_w} S_m = \frac{M_i S_i}{\rho_{mw}} + \frac{M_w S_w}{\rho_w}$$

$$\Delta S = (S_m - S_w) \sim \frac{M_i}{M_i + M_w}\sigma$$

where σ is the difference between salinity of sea ice and seawater (typically ~28), and ρ_{mw} is the density of meltwater, $\rho_{mw} \sim \rho_w$. The temperature of the mixture T_m is given by

$$T_m C_w (M_i + M_w) = T_i C_i M_i + T_w C_w M_w - M_i L$$

$$\Delta T = T_m - T_w = \frac{M_i}{M_i + M_w}\left(T_i \frac{C_i}{C_w} - T_w - \frac{L}{C_w}\right)$$

Then since

$$\left(T_i \frac{C_i}{C_w} - T_w\right) << L/C_w$$

$$\frac{\Delta T}{\Delta S} \sim \frac{-L}{C_w \sigma} \tag{1}$$

where C_w and C_i are the specific heats of seawater and ice. It should be noted at this stage that this equation applies only to the case when heat conduction through the ice is negligible and there is negligible warming of the ice by the underlying water. As such, the equation represents a limiting case of the changes in T and S that result simply from mixing sea ice and water which is above its freezing point. The implications of these restrictions will be discussed below.

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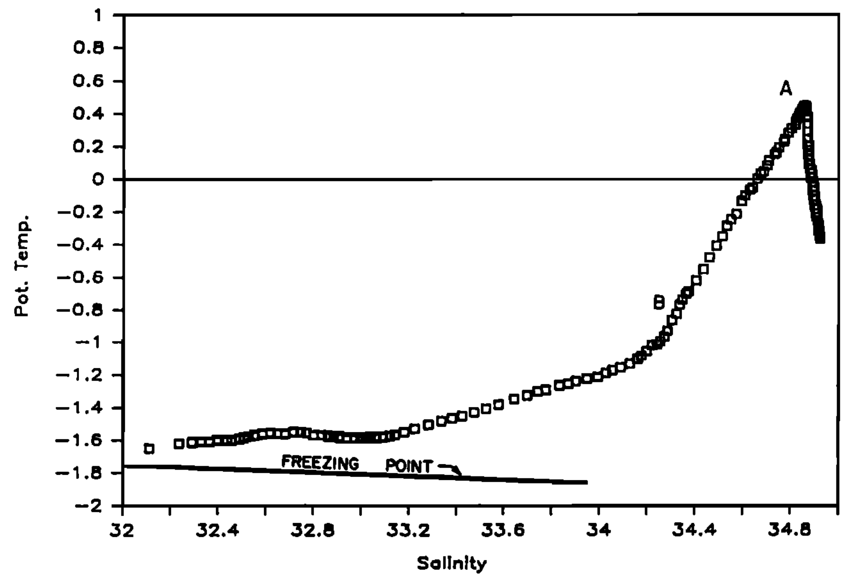


Fig. 1. Temperature-salinity relationship at CESAR station in the Arctic Ocean (R. Perkin and E. L. Lewis, unpublished conductivity-temperature-depth data, 1983).

For convenience of representation on a *T-S* diagram, the ice will be treated as a hypothetical end member of a mixing series with a temperature given by L/C_w. In a system where interaction occurs between a supply of Atlantic water and ice, the products of mixing have properties that lie along the mixing line shown in Figure 2, the ice properties being in this case $L = 2.67 \times 10^5$ J kg^{-1}, $S = 7$. The result is a family of waters having temperature and salinity characteristics between those of the Atlantic water and the point where the mixing line intersects the freezing point line. This point of intersection represents the product of bringing water above its freezing point into contact with a surplus of ice; all the sensible heat of the water is then utilised and an equilibrium system of ice and freezing point water results.

DISCUSSION

In Figure 3, *T-S* data for a number of stations for western Arctic locations (Figure 4) are reproduced; included are straight lines denoting the properties that would result from interaction of the Atlantic layer with ice under the restricted conditions specified above. It is seen that the hypothetical process would yield *T* and *S*) properties almost identical to those observed.

Data have been taken for a number of western Arctic stations, and straight lines have been fitted to *T-S* data lying between the Atlantic *T* maximum and the inflection point (at salinity of ~34.3). In Figure 5 these lines have been extrapolated into the region of the *T-S* plot occupied by *(L/C)-S* values of sea ice having salinities between 3 and 11 and temperatures of -1.6° to -2°C [*Ono*, 1967]. The numbers of data points used for linear regressions are as follows: CESAR, 62; AIWEX, 25; T3, 27; LOREX, 16. Figure 5 demonstrates that the agreement between observed *T-S* properties of waters in the Arctic thermocline and those predicted to result from interaction of "warm" waters with sea ice does not depend critically on the selection of the salinity, temperature, and hence latent heat of sea-ice.

In the case of stations lying close to the inflow of Atlantic water to the Arctic Ocean (e.g., Fram Strait) it is found that the thermocline $\Delta T/\Delta S$ slope is steeper than predicted by equation (1). *Perkin and Lewis* [1984] have accounted for the development of *T-S* relationships in the vicinity of Svalbard in terms of mixing between the water masses entering and leaving the Arctic Ocean. At these stations the Atlantic layer is relatively warm (~2.5°C), with its temperature falling rapidly during its transit along the Siberian continental margin. However, such stations are not representative of the Arctic Ocean as a whole.

At this point, some consideration should be given to the conditions that were attached to the validity of equation (1). Under what circumstances would the conductive heat flux through the ice be negligible compared with the melting

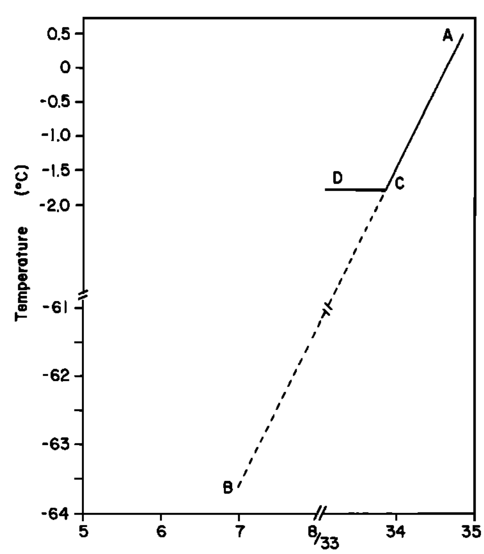


Fig. 2. *T-S* plot showing the mixing line AB, having end members of Atlantic water and sea ice ($S = 7$, $L = 2.67 \times 10^5$ J kg^{-1}). Line CD represents the freezing point as a function of salinity.

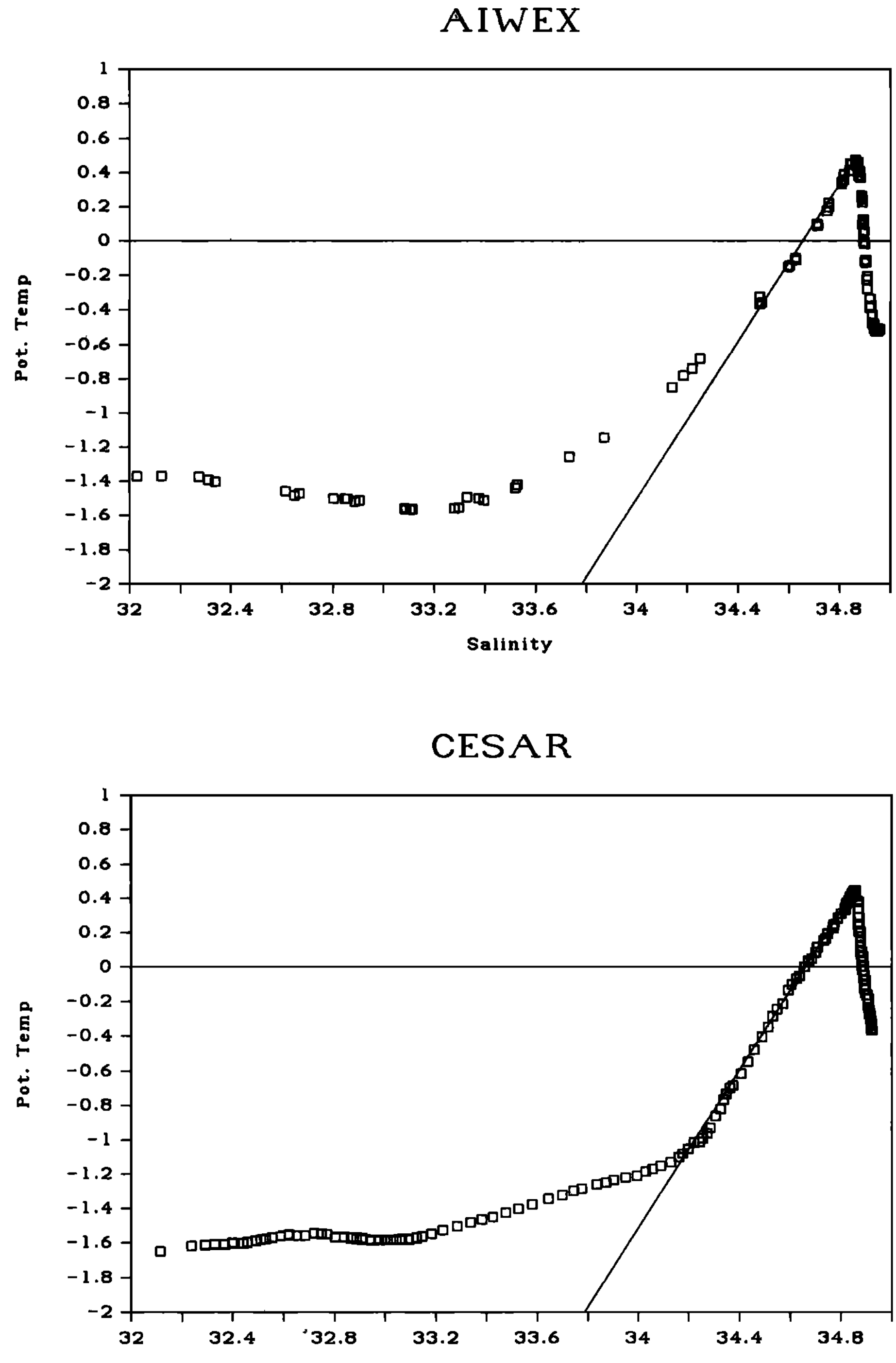


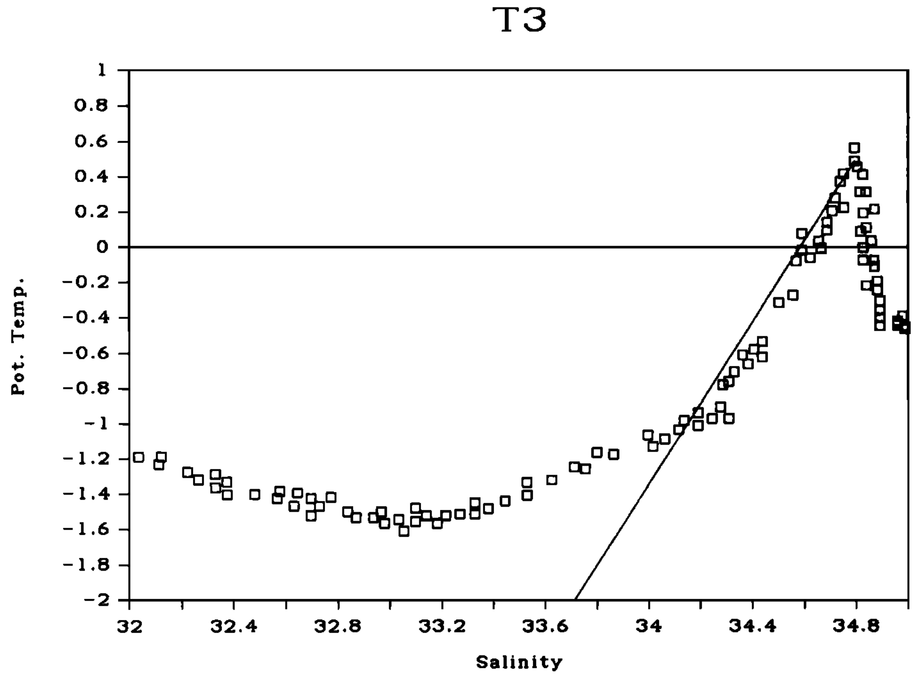
Fig. 3. *T-S* plots from a number of Arctic Ocean vertical profiles. Shown are predicted properties resulting from interaction of Atlantic layer with sea ice (S=7, L = 2.67 x 10^5 J kg^{-1}). CESAR data from R. Perkin and E. L. Lewis (unpublished data, 1983); Beaufort Sea Station 31 data are from H. Melling (unpublished data, 1981). T3 data are digitized from *Kinney et al.* [1970]. AIWEX data are from J. Swift (unpublished data, 1985).

rate? *Maykut* [1978] estimates that the seasonally varying conductive heat flux through sea ice of thickness 3 m and 0.8 m will have a wintertime maximum of 15 and 75 W m^{-2}, respectively. Melting rates that involved latent heat fluxes 1 order of magnitude greater than these conductive fluxes would be 5 cm d^{-1} for 3-m ice and 23 cm d^{-1} for 0.8-m ice.

An estimate can be made of the conditions that would be required in the Arctic Ocean to melt ice at a rate of 5 cm d^{-1}, which is equivalent to a heat flux of 140 J m^{-2} s^{-1}. Using the relationship [*Morison et al.*, 1987]

$$Q = C_h C \rho_w U_* \Delta T J m^{-2} s^{-1} \qquad (2)$$

where Q is the heat flux; C_h, the heat transfer coefficient (relative to friction velocity), is ~0.004 [*Morison et al.*, 1987]; C is the specific heat of water; U_* is the friction



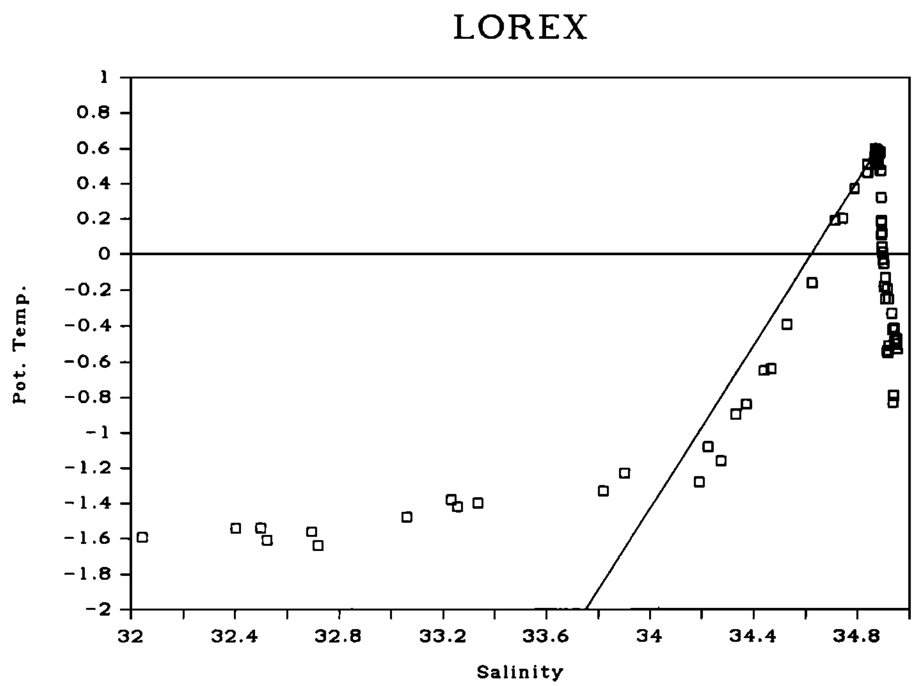


Fig. 3. (continued)

velocity ($= VC_d{}^{0.5}$); and the drag coefficient, C_d (relative to square of relative water velocity at 30 m) is taken as 0.006 [*Morison et al.*, 1987]. If the value of ΔT is taken as 0.5°C, we find from equation (2) that the required current velocity is ~0.2 m s^{-1}. An increase in ΔT would proportionately reduce the necessary current velocity. Such conditions need occur only periodically and do not necessarily require the ice cover to show any net decrease in thickness over considerably longer periods of time. (It might be noted that in the case of the thick ice of pressure ridges, any melting due to contact with "warm" water would have negligible associated conductive heat flux and would involve a progressive thinning of such ice.)

While this suggests that the process could occur at a significant rate, it is clear that a sufficiently vigorous mixing regime must exist at the surface to permit the process to proceed; in the absence of such mixing, the production of the cooler, slightly less saline water might tend to induce stable density stratification and so decrease the flux of sensible heat to the ice-water interface, although *Josberger* [1983] has argued that the stratification would not be enough to inhibit the melting in the marginal ice zone.

Clearly, if the melting were occurring at seasons when the air temperature were closer to freezing point (rather than the approximately -30°C temperature used in the above estimate), the conductive heat flux would approach zero.

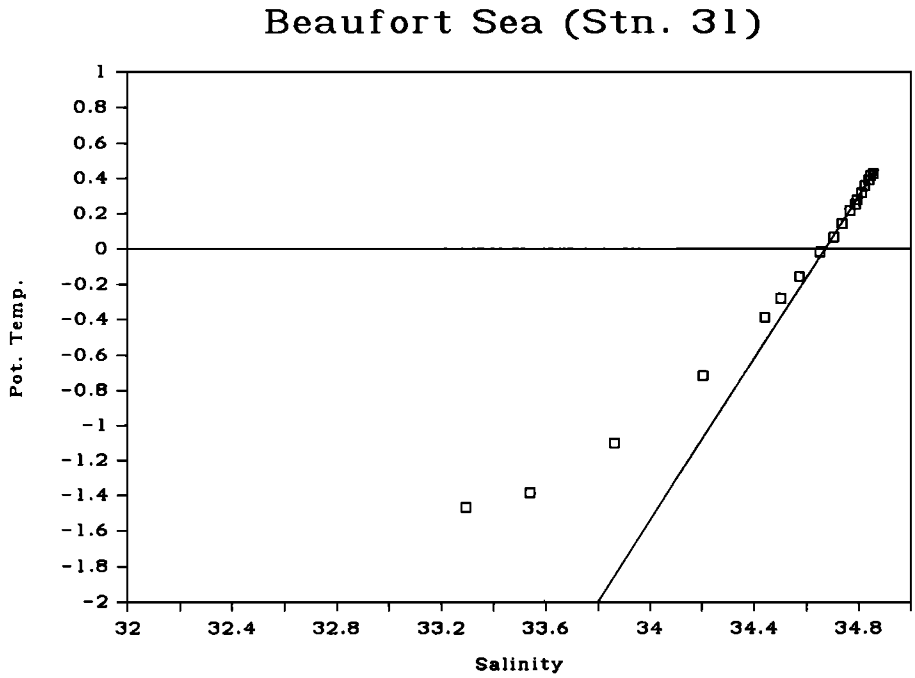


Fig. 3. (continued)

Josberger [1987] reports an ablation rate of 6 cm d^{-1} for ice in water which was $< 1°C$ above freezing point when the relative speed was 10 cm s^{-1}. A melt rate as high as 70 cm d^{-1} is reported by the same author when the water was 3°C above freezing and the relative velocity varied between 20 and 30 cm s^{-1}. These observations were made in the marginal ice zone, and although it would appear that this region could at times produce waters of the type being considered, the situation is complicated by the possibility of substantial heat exchange with the atmosphere. Indeed,

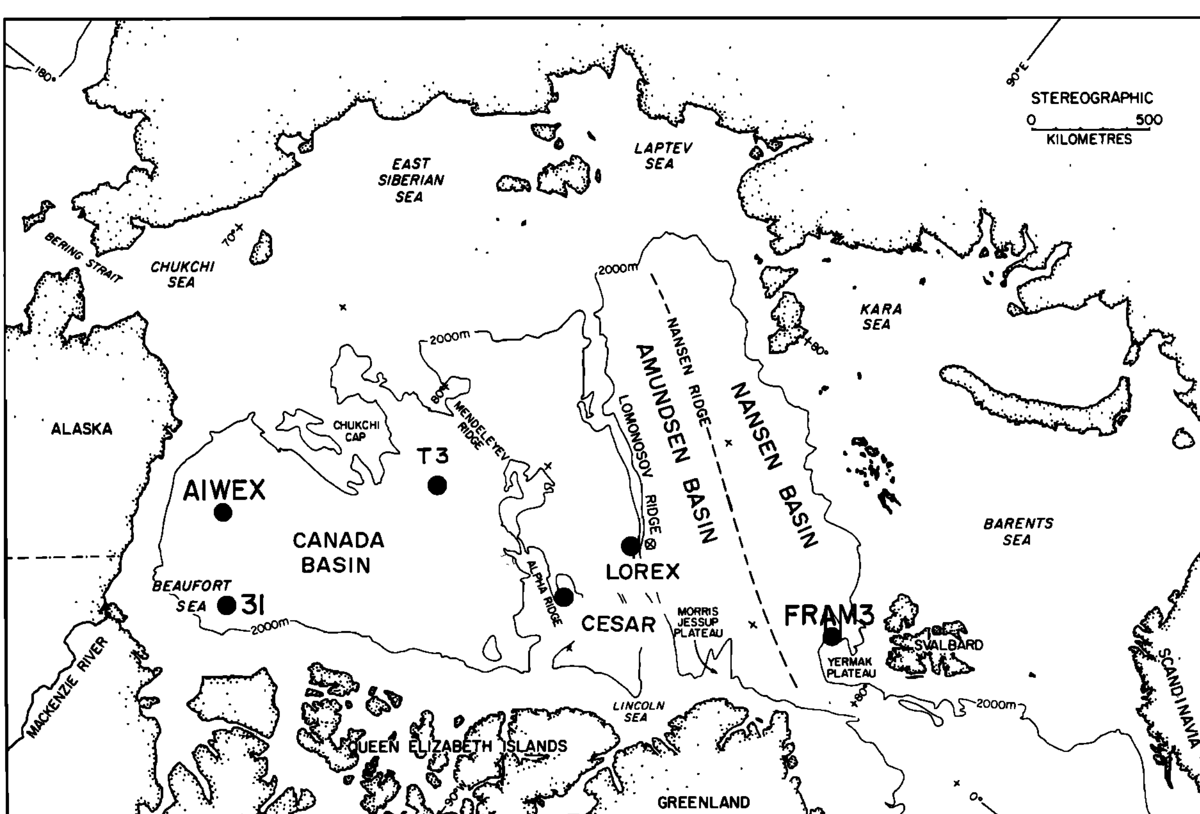


Fig. 4. Location of oceanographic stations.

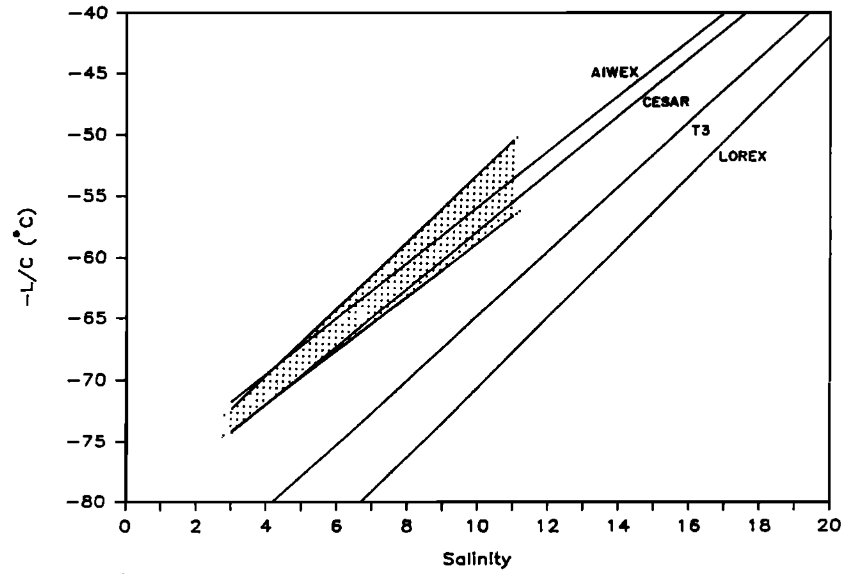


Fig. 5. Extrapolation of linear regression to thermocline *T-S* data for various stations into region of (L/C) and *S* properties of sea ice.

it is beyond the scope of this paper to identify areas of production of the Arctic intermediate waters having the specific *T-S* properties to which we have referred.

When the condition of negligible conductive heat flux is not met, a larger volume of water is required to melt a given amount of ice, and the slope of the *T-S* relationship will be steeper than is indicated by equation (1). This could account for the data of the T3 and LOREX stations (Figure 3) lying to the right of (or beneath) the ice-water mixing line, heat transfer to the atmosphere having produced additional cooling of the waters.

It is important to stress that the ideas expressed here do not require that water from the temperature maximum be brought to the ice boundary. What is required is that water above its freezing point should sporadically interact with the ice surface. Such waters commonly are observed in the western Arctic to lie at depths below ~150 m; the resulting cooled waters are presumed to mix with underlying warmer waters so effectively transporting heat from the Atlantic core.

An implication of these ideas is that there is the potential for rapid transfer of heat from the Atlantic layer over short time periods. The short-term heat fluxes from the water column can be very much higher than that associated with average ice accumulation rates: ice is then acting as a heat buffer, the freezing process providing a constant and relatively steady flux of heat to the atmosphere while occasionally accepting a high heat flux from upwelled warmer waters.

There is no indication in the literature, as far as we are aware, that any other investigator has recognised the remarkable correspondence between the observed thermocline *T-S* slope of Arctic Ocean waters with that predicted to result from the involvement of melting ice. We therefore argue that the true surface layer of the Arctic Ocean, that is, ice, has been overlooked as a possible contributor to the *T-S* properties of the upper water column. It seems reasonable to suppose, however, that certain of the transport mechanisms that earlier authors have sought to account for the properties of the cold halocline could also serve to produce the deeper thermocline waters by the mechanism suggested here. The general transport mechanisms of halocline water formation to which we refer are the flow of Atlantic waters to shallow depths via submarine canyons located in the continental shelves [*Coachman and Barnes*, 1962] and upwelling at shelf breaks [*Aagaard et al.*, 1981]. As originally envisaged, these mechanisms bring saline water above its freezing point to the surface where, through heat loss to the atmosphere, freezing-point waters are produced; these have properties indicated by the line CD in Figure 2. We point out that if the upwelled waters from the Atlantic layer are cooled by melting some ice they will then have properties that lie on the line AC. It should be noted that existence of two adjacent water types (AC and CD) will cause the development of a new mixing product. It is to be expected that decay of the boundary between the two water types should increase with geographic distance from the source of the ice-cooled waters. Examination of the *T-S* relationships of Figure 3 suggests that of the stations we can consider, AIWEX is most distant from such a source, and the LOREX station on the eastern side of the Lomonosov Ridge is the closest.

In their attempt to model the vertical *T-S* structure of the water column of the Arctic Ocean, *Killworth and Smith* [1984] found insufficient cooling by mixing between Atlantic waters and cold waters lying above and below to counteract the effects of the warm Atlantic inflow. They questioned whether the poor simulation below 200 m might be due to the omission of a key piece of physics; we suggest that the process outlined here may provide the key.

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