

## Measurements of the electrical conductivity of interstitial water in subsea permafrost

W.D. HARRISON AND T.E. OSTERKAMP

*Geophysical Institute, University of Alaska, Fairbanks, Alaska 99701, USA*

Interstitial water samples have been obtained from the thawed layer beneath the sea bed at Prudhoe Bay, Alaska, using a probe method. The electrical conductivities of the water samples, and therefore the salinities, are about 25 per cent higher than those of normal sea water. This difference may be due to a density-filtering process caused by convection of the interstitial water. The high salinity causes the phase boundary temperature at the bottom of the thawed layer, where ice-bearing permafrost exists, to be lower than the freezing point of normal sea water. A rather uniform value of  $-2.4^{\circ}\text{C}$ , corresponding to a salinity of about 43 per mille, is found out to 3.5 km from shore. A downward salt flux exists at the bottom of the thawed layer, and the electrical conductivity of the interstitial water at one site shows evidence for a thin boundary layer there, in which the salt transport regime seems to change from convective to diffusive. Above this layer, the salinity gradients are low, as would be expected in a well-developed convective regime. A characteristic interstitial water speed at a site 700 m from shore appears to be of the order of a few tenths of a metre per year.

On a prélevé des échantillons d'eau interstitielle dans la couche dégelée du fond sous-marin de la baie de Prudhoe en Alaska, à l'aide d'une sonde. La conductivité, et par suite la salinité des échantillons d'eau, étaient de 25 pour cent supérieures aux valeurs caractérisant l'eau de mer normale; il s'est peut-être produit un filtrage suivant les densités, causé par la convection de l'eau interstitielle. La salinité élevée a abaissé la température de l'interface entre les phases à la base de la couche dégelée, où se trouve le pergélisol riche en glace, au-dessous du point de congélation normal de l'eau de mer. A 3,5 km du rivage, on a relevé une température plutôt uniforme de  $-2,4^{\circ}\text{C}$ , correspondant à une salinité d'environ 43 pour mille. A la base de la couche dégelée, on a observé un flux salin descendant, et la conductivité électrique de l'eau interstitielle en un certain site indique l'existence d'une mince couche limitrophe, où le régime de transport des solutions salines change (diffusion au lieu de convection). Au-dessus de cette couche, les gradients de salinité sont faibles, comme habituellement dans un régime de convection stabilisé. En un site placé à 700 m du rivage, la vitesse caractéristique de l'eau interstitielle semble être de quelques dixièmes de mètres par an.

Proc. 4th Can. Permafrost Conf. (1982)

### Introduction

In the last decade, several research programs have demonstrated that permafrost underlies much of the shelf area of the Arctic Ocean. The authors' effort has been part of a co-ordinated program to study the permafrost underlying the shelves of arctic and sub-arctic Alaska. At present, it has two main objectives: First, the mapping of permafrost conditions over remote areas, with reconnaissance techniques using portable driving and jet drilling equipment transported by helicopter or snow machine; and secondly, the study of the heat and mass transport processes that determine the properties and evolution of the permafrost. The techniques and similar ones used in Canada, do not replace conventional drilling and soil sampling programs, but allow a great deal of data to be collected relatively cheaply, and in certain cases, data that are more difficult to get by conventional methods. Such data include hydraulic conductivity, *in situ* thermal properties, and interstitial water samples from permeable sediments. The probe-driving techniques used to obtain the data described here have been described by Harrison and Osterkamp (1981) and by Osterkamp and Harrison (1981, 1982).

They are used to set expendable plastic tubing into the sea bed for temperature and thermal property measurements, or to obtain interstitial water samples and hydraulic conductivity data with a special sampling probe. The probe method is rather efficient because it is not necessary to pull the probe between samples.

A study of the electrical conductivity of interstitial water, a measure of its salt content, is described in this paper. A summary of some of the related temperature measurements is given in a companion paper (Osterkamp and Harrison 1982). Salt is a fact of life in marine environments, and, because of its control of phase relationships, it must be taken into account in efforts to understand the evolution and properties of subsea permafrost. Thus, a new variable, the interstitial water salinity  $S$ , enters on an equal footing with the temperature  $T$ . In fact, there is a striking similarity in the transport equations for heat and salt, and, in some situations, the rate of evolution of subsea permafrost can be thought of as being controlled by the transport of salt rather than heat. However, there is a strong coupling, basically because  $S$  and  $T$  are related by phase relationships at ice-water boundaries (Harrison and Osterkamp 1978).

### Setting

The data described here were obtained on the west side of Prudhoe Bay, Alaska, along a study line bearing about N31°E from the North Prudhoe Bay State Number One well and crossing Reindeer Island, about 14 km distant. The line crosses the shore near the ARCO west dock (Figure 1). Drilling, sampling, probing, and temperature measurement programs have been conducted in this area by several investigators since 1975. A review of the literature is given by Sellmann (1980). In what follows, the University of Alaska holes are designated by their distance from shore in metres. Those drilled by the U.S. Army Cold Regions Research and Engineering Laboratory, and the U.S. Geological Survey program, are designated by CRREL, followed by the hole name. The lithology, where the data were obtained, consists mainly of outwash and alluvial sediments, probably throughout the thickness of the permafrost, which is about 560 m on-shore at the North Prudhoe Bay State Number One well (Osterkamp and Payne 1981). Several metres of recent silty marine sediments overlie this material off-shore. The hydraulic conductivity in the thawed layer beneath the sea bed, which is determined in the course of the interstitial water sampling, is typically 1 to 10 m/a.

Within a few kilometres from shore, the permafrost has been inundated for a time that is determined by the rate of shore-line erosion; eustatic changes in sea level can probably be ignored. The erosion of Alaska's coasts has been studied by many observers (see Hopkins and Hartz 1978). In the study area, there seems to be some disagreement about the rate of erosion of the shore-line, Cannon (1979) and Lewellen (1977) giving values of 3.7 and 1.5 m/a, respectively, averaged over about the last 25 years. However, there is some lateral variability, and Cannon suggests a spatial average for the west side of Prud-

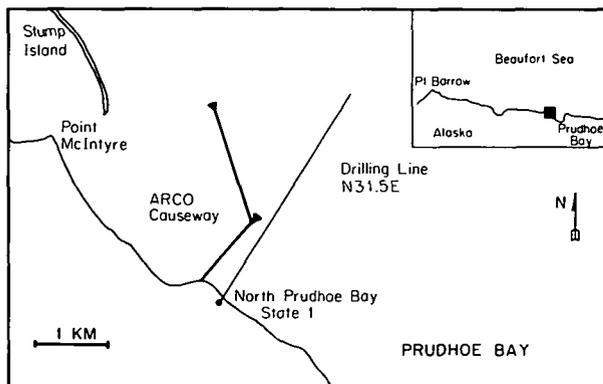


FIGURE 1. Location map of study area.

hoe Bay of 2.8 m/a. Interpretation of borehole thermal data by Lachenbruch and Marshall (1977) suggests a rate of 1.0 m/a averaged over the last few thousand years, while the authors find a rate of somewhat less than 1 m/a averaged over the last few hundred. The authors have adopted 1 m/a as a representative value, realizing the uncertainty and that the rate may not be constant.

### Results

Data show the electrical conductivity of the interstitial water from the sea bed through the underlying thawed layer to the ice-bonded permafrost beneath it (Figures 2, 3, and 4). Data tabulations are available (Osterkamp and Harrison 1980). These data were obtained from two sites at 438 and about 700 m from shore respectively. Hole 701-1980 could not be located exactly with respect to the similar 1979 holes

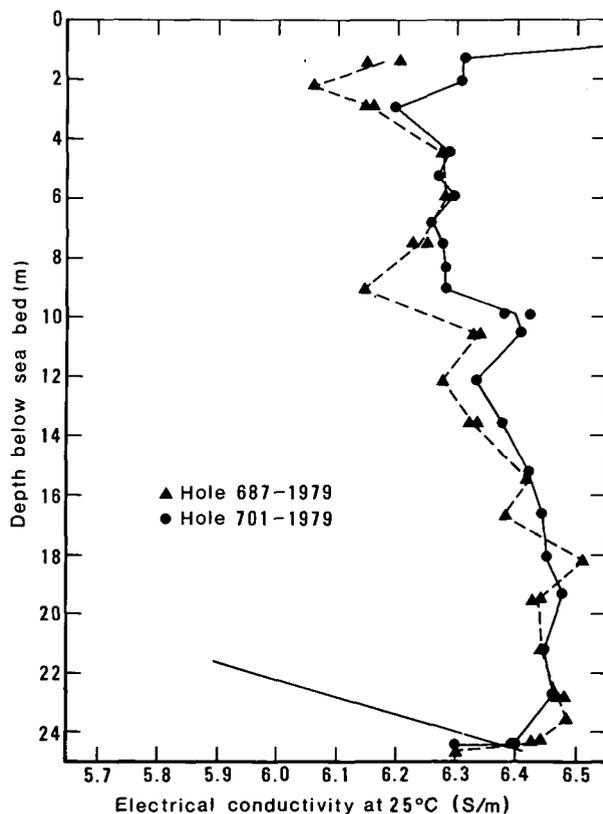


FIGURE 2. Electrical conductivity of the interstitial water through the thawed layer at the 700-m site in 1979. The thawed layer is about 24.5 m thick. The hole number is the distance from shore in metres. In 1979, the probe sampling length was 0.1 m. The diagonal straight line represents the interstitial water conductivity that would exist if the water were at its freezing point as estimated from the temperature data in Figure 7.

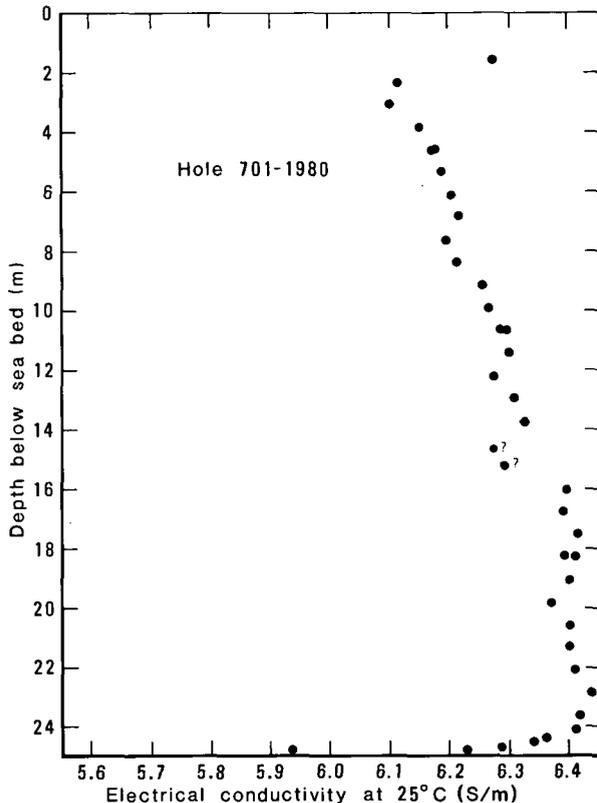


FIGURE 3. Electrical conductivity at the 700-m site in 1980. The 1979 and 1980 sites could be displaced 10 m laterally.

because one of the 1979 reference marks was destroyed; the 1980 holes could be displaced 10 m laterally from the others.

The electrical conductivity of the interstitial water samples at 25°C was determined in the laboratory to a precision and accuracy of half a per cent or better. A study of the chemical composition of the interstitial water in this vicinity by Page and Iskandar (1978)

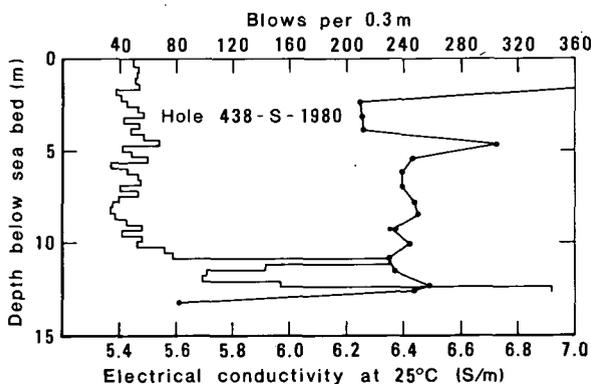


FIGURE 4. Electrical conductivity (right) and blow count data (left) through the thawed layer at the 438-m site.

indicated that the major ions have ratios characteristic of normal sea water. However, this interstitial water is typically 25 per cent saltier than normal sea water, which has an electrical conductivity, at 25°C, of 5.307 S/m ( $S = \text{ohm}^{-1}$ ). Because of its similarity to sea water, the salinity and freezing point temperature of our samples can be estimated by standard methods (UNESCO 1966; Doherty and Kester 1974), but an extrapolation outside the range of measured values is necessary.

### Discussion of Data

The interstitial water data obtained in 1979 and 1980, and its relationship to other data obtained from soil samples in 1975, 1976, and 1977 (Iskandar *et al.* 1978; Page and Iskandar 1978), are discussed in this section.

#### Conductivity Near the Sea Bed

All electrical conductivity profiles obtained by the probe method show a marked increase near the sea bed. Although this behavior plots off the scales of Figures 2, 3, and 4, more information is available (Osterkamp and Harrison 1980). The effect is caused by seasonal partial freezing of the sea bed, and concentration of the salt in the remaining liquid phase. The probe method samples this liquid, and not the bulk water (Harrison and Osterkamp 1981). The partial freezing extended to a depth of roughly two metres below the sea bed in 1980 at both sites. This seasonally active layer beneath the sea bed, as well as an irregularly varying bulk water conductivity in it, which is probably also related to concentration by seasonal freezing, has been noticed by all the investigators who have worked in this area.

#### Conductivity Below the Active Layer

Except for some notable details, the conductivities at the off-shore sites (see Figures 2, 3, and 4) are similar in magnitude, roughly 6.3 S/m, despite the fact that the 438-m site is in shallower water (1.5 m compared with 1.8 m in 1979 and 1.7 m in 1980 at the 700-m site), and therefore subject to higher concentrations of late winter salt at the sea bed due to downward freezing of the sea ice. A five per cent increase in conductivity from top to bottom of the thawed layer is evident at the 700-m site (see Figures 2 and 3). Perhaps a more precise interpretation is that the conductivity increases to a depth of roughly 15 m below the sea bed, and then is roughly constant down to some fraction of a metre above the frozen boundary. The structure evident in the conductivity in holes 687 and 701-1979 (see Figure 2) tends to reproduce in both holes, and is therefore probably real.

An outstanding question is why the conductivity of the interstitial water and therefore the salinity, are typically 25 per cent higher than those of normal sea water. Much of the salt must have entered the sediments after they were inundated by shore-line erosion because the bulk salt content of shallow samples obtained nearby on-shore (holes 0, -225, -226, and CRREL PB-9) is relatively low, although some evidence exists for high salt contents in the upper 180 m of sediments in well logs (Osterkamp and Payne 1981). The sharp thawed-frozen interface described below also indicates present-day salt transport, as discussed later. A partial explanation of how this may take place was provided by the authors (Harrison and Osterkamp 1978), who showed that the generation of relatively buoyant water by thawing permafrost can lead to gravitational instability and convection of the interstitial water. This kind of instability can give rise to rapid salt transport from the overlying sea water into the thawing material. Whether it exists, depends to a large extent on the permeability of the sediments; the prediction that it should exist in this area is based on our measurements of the hydraulic conductivity. At first sight, it might seem that this efficient salt transport mechanism might lead to an interstitial water salinity the same as the mean annual value of the overlying sea water. This is not the case, because the convective process should act as a density filter, permitting heavy, salty water generated by downward freezing sea ice to sink into the sediments, and excluding the lighter water present at the sea bed at other times. This is probably why the interstitial water is saltier than normal sea water. It is possible that the effect may also be able to explain the slight increase in salinity with depth at the 700-m site, since the heavier plumes of salty water may sink through the lighter water and give rise to the observed stratification.

The electrical conductivity profiles presented here seem to show less variation with depth than others obtained along the same study line at distances from shore of 481,  $\approx$  870 (CRREL PB-6), 3370, and  $\approx$  3500 (CRREL PB-7) m. At least some of the variation must be real. However, the interstitial water from these holes was extracted from soil samples obtained with drill rigs, and it seems possible that in a few cases contamination may have occurred (see Harrison and Osterkamp 1981). Nevertheless, in these holes the interstitial water is considerably saltier than normal sea water, as it is in hole 438-S and at the 700-m site. It is therefore not surprising, because of phase equilibrium, that the phase boundary at the bottom of the thawed layer in all of these holes is colder than the freezing point of normal sea water. What is surprising is that the observed temperature,

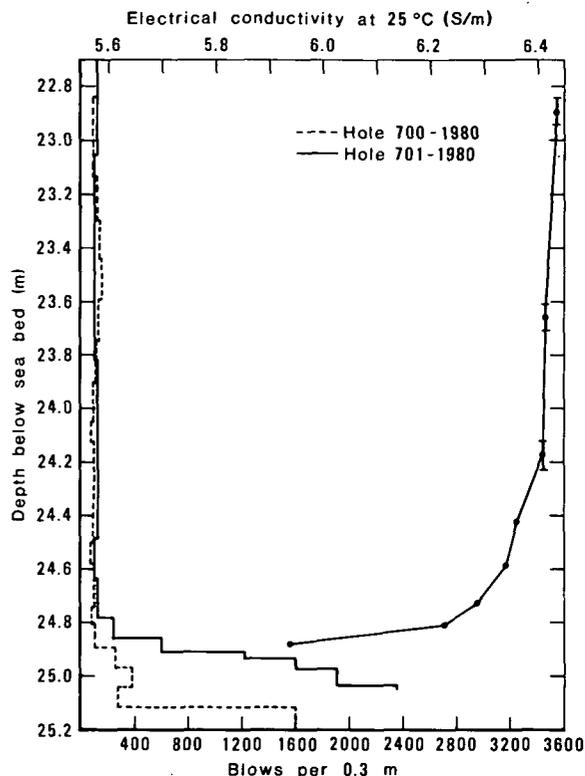


FIGURE 5. Expanded view of electrical conductivity (right) and blow count data (left) near the bottom of the thawed layer at the 700-m site in 1980. The probe sampling length was 0.01 m for the lowest five points and 0.1 m above. The water samples were obtained from hole 701, and the conductivity should be compared to the blow count in the same hole. The blow count in the adjacent hole 700 is also shown.

about  $-2.4^{\circ}\text{C}$ , is about the same in all of these holes. This means that the salinity at the phase boundary is uniform as well, at about 43 per mille, even though there is not known to be any seasonal concentration of salt beneath the sea ice at the 3370- and 3500-m holes. Unfortunately, the authors do not understand the reasons for this apparent uniformity, or what determines this particular value. Several holes farther from shore have been sampled in the study area as well (CRREL PB-3 at 6.62 km, PB-8 at 12.1 km, and PB-2 at 17.0 km), and these are different in that the interstitial water salinity, while showing some variations, is more characteristic of normal sea water, as are the phase boundary temperatures at the bottom of the thawed layer.

#### Conditions Near the Phase Boundary

A marked decrease in conductivity is evident in the vicinity of the ice-bonded boundary, which occurs at the 700- and 438-m sites at depths of about 25 and 12 to 13 m respectively (see Figures 2, 3, and 4). It is

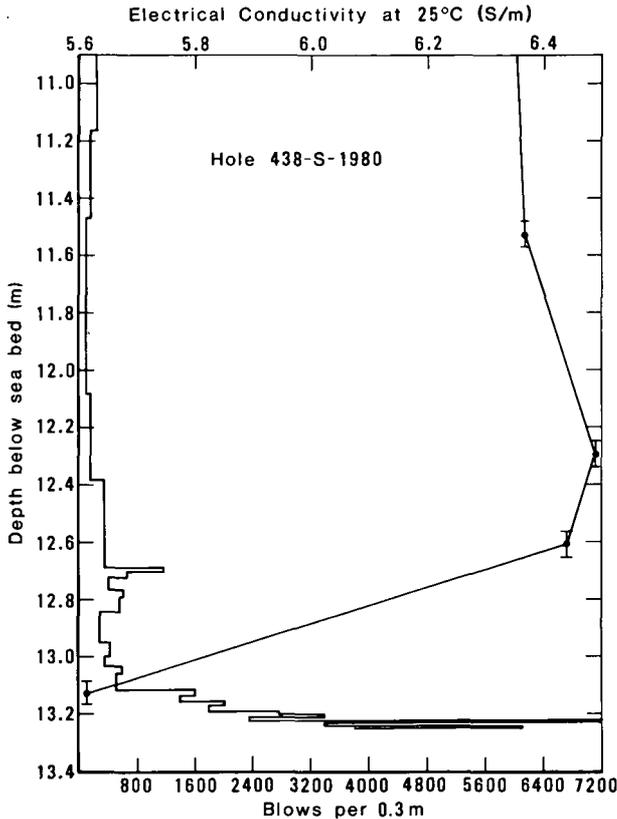


FIGURE 6. Expanded view of electrical conductivity (*right*) and blow count data (*left*) near the bottom of the thawed layer at the 438-m site. The bars indicate the probe sampling length of 0.1 m.

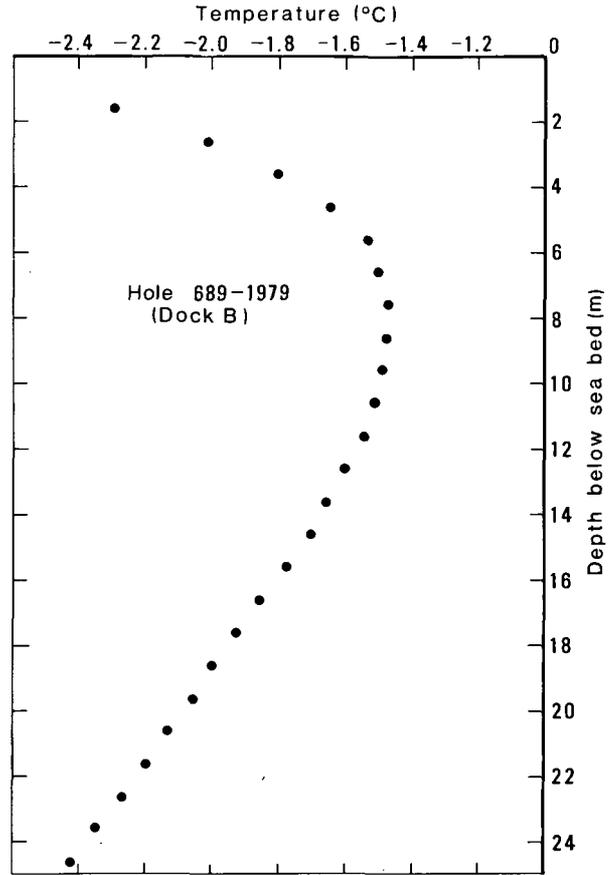


FIGURE 7. Equilibrium temperature at the 700-m site in 1979.

important to determine exactly where this decrease takes place. For this purpose, an expanded view is given of the conductivity data near the bottom of hole 701-1980, together with the blow count data from the driving of the probe, both in this hole and in the adjacent hole 700-1980 (Figure 5). (The blow count is the number of hammer blows to drive the probe 1 foot or  $\approx 0.3$  m). Similar data for hole 438-S are given (Figure 6); and blow count data with less detail near the boundary are also given (see Figure 4). Temperature data for hole 689-1979 are shown (Figure 7). The diagonal straight line near the bottom of holes 687 and 701-1979 (see Figure 2) represents the interstitial water conductivity that would exist if the water were at its freezing point, as estimated from the temperature data (see Figure 7). The lowest part of this line gives an independent estimate of the conductivity at the boundary, but note that this requires extrapolation beyond the range of conductivity-salinity-freezing point measurements. A similar use of temperature data at the 438-m site (Osterkamp and Har-

risson 1982) has not been made because of some uncertainty in its absolute accuracy. At any rate, the temperature-implied conductivity at the boundary in hole 438-S is similar to that in Figure 2. An examination of these data, and of more details given by the authors (Osterkamp and Harrison 1980) leads to the following conclusions at the 700-m site:

1. Above the ice-bonded boundary defined by the blow count data, the measured conductivities are higher than those estimated to exist if the water were at its freezing point. Therefore no ice is present there.
2. The mechanically ice-bonded boundary is well defined, to within a thickness of roughly 0.2 m.
3. There is a marked decrease in conductivity as the probe approaches and penetrates the boundary.
4. The conductivity begins to decrease in hole 701-1980 before the boundary is reached. The same also seems to be true in holes 687 and 700-1979, although the deepest measurements in these holes were made after the boundary had been encountered.

At first glance, the situation at the 438-m site seems

similar (see Figure 6). However, one does not know exactly where or how well defined the boundary is there, because the temperature data are not as accurate, and because the blow counts below about 12.5 m are higher than one would normally expect from ice-free material in this region, even though there is a well-defined sharp increase at about 13.1 m. Consequently, conclusions 1, 2, and 4 may, or may not, apply in hole 438-S, although conclusion 3 does.

Conclusion 1 is the basis for using the terms "ice-bonded" and "phase" boundary interchangeably in this area. Some care must be exercised, because ice-bearing material is not necessarily ice-bonded. For example, the presence of seasonal ice in the soil near the sea bed means that it is sometimes ice-bearing, but often there is little evidence of mechanical bonding. Conclusion 3 is expected if ice exists in finite amounts below the boundary, because it can be shown that enough energy is dissipated at the probe tip to melt some of the ice when the driving is hard, thereby diluting any liquid solution that may be present in the ice. In this situation, the probe probably samples neither the liquid nor the bulk water salinity but something in between. Therefore the bulk water salinity is probably much lower than indicated by the deepest points in the data. The marked decrease in bulk water salinity below the boundary is due to the presence of ice, and has been observed before in this region in hole 481-1975. This is a different effect from the small decrease in salinity above the phase boundary at the 700-m site (conclusion 4).

The discontinuity in bulk water salinity across the moving phase boundary implies a finite salt flux there, which can be estimated from the salinity at the boundary (43.3 per mille), the initial bulk water salinity of the permafrost, and the thaw rate. The thaw rate is estimated as follows. A parabola of the form  $Y = a \sqrt{x-x_0}$  is a good fit (standard deviation 1 or 2 m) to the thawed-frozen boundary as defined by all the drill holes out to 3.5 km from shore. Here,  $Y$  is the thawed layer thickness measured from the sea bed,  $x$  is the distance from shore, and  $(a, x_0) = (1.147 \text{ m}^{1/2}, 276 \text{ m})$ . If the shore-line retreat rate,  $V$ , can be considered constant, which is uncertain,  $x = Vt$ , where  $t$  is the time that the permafrost has been

inundated. This means that

$$[1] \quad Y = a' \sqrt{t-t_0},$$

where  $(a', t_0) = (\sqrt{V}a, x_0/V)$ , which is similar to the Stefan problem result. From equation 1 the thaw rate  $dY/dt$  is given by

$$[2] \quad \frac{dY}{dt} = \frac{Y}{2(t-t_0)}.$$

A working value for the shore-line retreat rate  $V$  is 1 m/a as discussed earlier, which implies  $t_0 \approx 276$  a, and at the 700-m site,  $t \approx 700$  a. For the  $Y = 25$  m at this site, it follows that  $dY/dt \approx 0.029$  m/a. The salt flux  $Q$  is expressed in terms of the thaw rate by a phase boundary relationship that is the analogue of the Stefan boundary condition for heat:

$$[3a] \quad Q = (S_Y - S_i) \frac{dY}{dt}$$

or

$$[3b] \quad Q = (S_Y - S_i) \frac{Y}{2(t-t_0)}$$

by equation 2. Here,  $S_Y$  is the salinity an infinitesimal distance above the boundary (43.3 per mille) and  $S_i$  is the initial bulk water salinity below the phase boundary. The flux,  $Q$ , is expressed in terms of pore cross section area. Equations 3 assume that no salt flux penetrates the ice-bearing boundary, which may not be strictly true. Since  $S_i$  is unknown, two reasonable values are considered, 0 and 25 per mille; the latter is the value found in hole 481-1975 just below the phase boundary. The resulting fluxes at the 700-m site are given in Table 1.

#### Saline boundary layer

The sharp salinity gradient that seems to exist just above the phase boundary at the 700-m site (see conclusion 4) remains to be discussed. If the phase boundary is impermeable, a convective salt transport regime cannot extend all the way down to it, because the normal component of the interstitial water velocity field must go to zero there. This means that close to an impermeable boundary the salt transport regime should change to diffusive, and be characterized by a

TABLE 1. Salinity change across the boundary layer  $\Delta S$ , salt flux  $Q$ , strongly diffusive layer thickness  $B$ , and characteristic interstitial water speed  $v$  for two values of the bulk  $\text{H}_2\text{O}$  salinity before thawing  $S_i$

Hole	$\Delta S(‰)$	$Q(\text{kg}/\text{m}^2/\text{a})$	$S_i = 0$		$S_i = 25‰/\infty$		
			$B(\text{m})$	$v(\text{m}/\text{a})$	$Q$	$B$	$v$
687-1979	0.5(4)	1.3	$1.7 \times 10^{-3}$	2.3	0.55	$4.1 \times 10^{-3}$	1.0
701-1979							
701-1980							

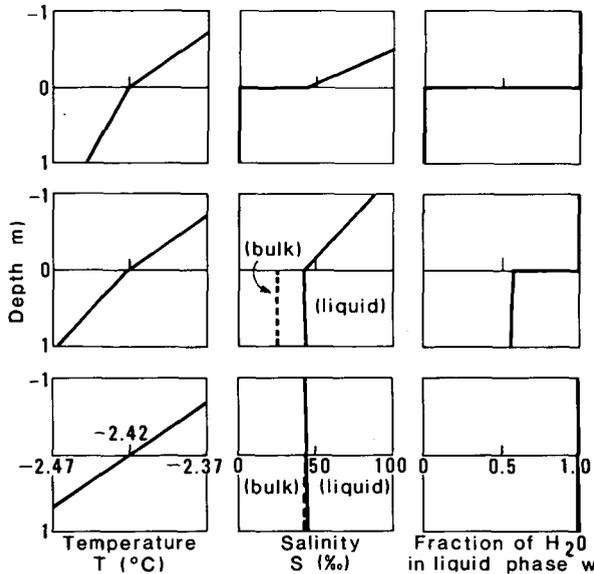


FIGURE 8. Three models of the behaviour of temperature, liquid water salinity, and the fraction of the water in the liquid phase at a phase boundary thawing at 0.01 m/a, where the depth scale is taken to be zero at the boundary. The broken line is the bulk  $H_2O$  salinity before thawing. It is 0, 25, and 43.3‰ in the top, centre, and bottom rows respectively.

large salinity gradient necessary to transport the salt diffusively to the melting ice. The layer of large gradient, and the layer of smaller, but still rapidly changing, gradient above it, will collectively be termed the saline boundary layer.

Three models of the behavior of temperature,  $T$ , interstitial water salinity,  $S$ , and fraction of water in the liquid phase,  $w$ , at an impermeable phase boundary are shown (Figure 8). These results apply only at the phase boundary itself, but are shown projected a metre above and below for clarity. They are obtained by applying the conditions that describe the conservation of heat and salt, and phase equilibrium at the boundary. The necessary thermal properties have been estimated by the methods summarized by Gold and Lachenbruch (1973), using parameters and methods similar to those in Harrison and Osterkamp (1978). These particular models assume that the thaw rate is 0.01 m/a, which is about a factor of 3 smaller than the value estimated earlier for the 700-m site. The difference in the three rows of Figure 8 is the initial bulk water salt content, which is 0, 25, and 43.3 per mille in the top, centre, and bottom rows, respectively. The value for normal sea water is 35 per mille, 25 per mille is the value measured in a sample from just below the phase boundary in hole 481-1975, and 43.3 per mille is the salinity of water in equilibrium with ice at the measured phase boundary tempera-

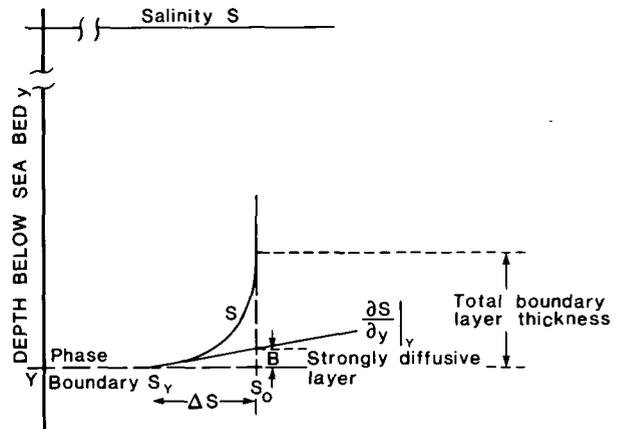


FIGURE 9. Idealized description of the behaviour of salinity in the vicinity of the boundary layer at the 700-m site. The symbols are defined on the figure. The line labelled  $\frac{\partial S}{\partial y} \Big|_Y$  is tangent to the salinity curve at  $y = Y$ .

ture, about  $-2.42^\circ\text{C}$ , at the existing pressure. The bottom row describes the melting of salty ice when there is no appreciable salt transport. There are no appreciable salinity or liquid water content gradients at the boundary, a situation that is ruled out by the observations described earlier. Therefore, there should be a finite salt flux, and one might expect to find a large salinity, or conductivity, gradient at the boundary as predicted by the first two models, unless the effect is somewhat smeared out by permeability of the ice-bearing surface. This is probably the origin of the large gradient noted in conclusion 4.

The boundary layer can be discussed using terminology which describes the observations above the phase boundary at the 700-m site in idealized form (Figure 9). The salinity,  $S_0$ , just above the boundary layer can be estimated directly from Figures 2, 3, and 5; the salinity  $S_Y$  at the boundary is probably most accurately estimated from temperature data (such as in Figure 7). The difference  $\Delta S$ , the change in salinity across the boundary layer, is a useful parameter because several quantities can be expressed in terms of it. Values for  $\Delta S$  at the 700-m site (see Table 1) could easily be in error by a factor of two or more. Those from hole 687-1979 seem most reliable because  $S_Y$  could be estimated from temperature measurements made in hole 689-1979 (see Figure 7) only two metres away. The same value for  $S_Y$  was assumed for hole 701-1979.  $S_Y$  for hole 700-1980 was obtained from Figure 5 by assuming the boundary to be at the depth where the blow count began to increase, but it is recognized that the actual phase boundary could be either slightly above or below this point.

The thickness  $B$  of the strongly diffusive part of the boundary layer can be estimated as follows: Where

the salt transport is diffusive, as assumed at the phase boundary, the flux  $Q$  is given by

$$Q = -k_s \left. \frac{\partial S}{\partial y} \right|_y$$

where  $k_s$  is the molecular diffusivity of salt in the sediments (roughly  $0.004 \text{ m}^2/\text{a}$ , Harrison and Osterkamp 1978). From Figure 9,  $-\left. \frac{\partial S}{\partial y} \right|_y = \frac{\Delta S}{B}$ . Therefore

$$[4] \quad B = \frac{k_s \Delta S}{Q}$$

Notice that this is a measure only of the thickness of the strongly diffusive part of the boundary layer. Above it, there is a layer of intermediate gradient that probably marks a transition between a diffusive and a convective transport regime. Values of  $B$  are given (see Table 1) for the two values of  $Q$  estimated in the last section.

The order of magnitude of the vertical component,  $v$ , of the interstitial water velocity can also be estimated from  $\Delta S$ . Above the boundary layer, where the regime is convective, the salt flux is of order  $vS_0$ . If this is thought of as a downward flux, there must, somewhere, be an upward flux carried by water that has lost some of its salt. This is of the order of  $vSy$ , so the net flux  $Q$  is of order  $v \Delta S$  and

$$[5] \quad v \sim \frac{Q}{\Delta S}$$

Note that equations 4 and 5 imply

$$[6] \quad v \sim \frac{\kappa_s}{B}$$

which might have been guessed on dimensional grounds. Values of  $v$  from equation 5 are given (see Table 1) for the two values of  $Q$  estimated in the last section.

The boundary layer interpretation summarized in Table 1 suggests that a characteristic interstitial water speed,  $v$ , is in the order of  $1 \text{ m/a}$  at the 700-m site. That this is probably too high can be seen from another approach. The temperature profile of hole 700-1979 (see Figure 7) seems quite linear below the depth of seasonal cooling (roughly  $12 \text{ m}$ ). This suggests that it is primarily diffusive, so the Peclet number for heat transport  $vY/\kappa \ll 1$  (Harrison and Osterkamp 1978) where  $Y = 25 \text{ m}$  is the thickness of the thawed layer, and  $\kappa \approx 25 \text{ m}^2/\text{a}$  is the thermal diffusivity. This gives  $v \ll 1 \text{ m/a}$ , which suggests that the boundary layer estimate is too high. A lower limit on  $v$  can probably be taken to be the estimated thaw rate,  $0.03 \text{ m/a}$ ; this gives  $v \gg 0.03 \text{ m/a}$  at the 700-m

site. Combining these limits suggests that the order of magnitude of the interstitial water velocity is one tenth to a few tenths of a metre per year at the 700-m site. This implies, by equation 6, that the strongly diffusive layer thickness,  $B$ , is one hundredth or several hundredths of a metre and, by equation 5, that  $Q/\Delta S$  has been overestimated. If  $\Delta S$  were well determined by the data, this would indicate that the salt flux,  $Q$ , is actually smaller than estimated, which would correspond to a high initial salt content (25 per mille or more), or to a thaw rate appreciably less than  $0.03 \text{ m/a}$ ; in the latter case, a shore-line retreat rate appreciably less than  $1 \text{ m/a}$  would be implied. In fact, such conclusions are probably not justified at this stage, given the uncertainties in both the data and its interpretation. But the fact that reasonably good evidence for the existence of a boundary layer has been found is by itself important evidence for the existence of convection, the limitations on a quantitative interpretation notwithstanding.

### Summary

The interstitial water salinity in the thawed layer beneath the sea bed is about 25 per mille higher than that of normal sea water in the area studied. It is suggested that this is due to a density-filtering process caused by convection in the interstitial water. The high salinity causes the phase boundary temperature of the bottom of the thawed layer to be lower than the freezing point of normal sea water. A uniform value of  $-2.4^\circ\text{C}$ , corresponding to a salinity of about 43 per mille, is found in a line of holes out to  $3.5 \text{ km}$  from shore. Holes farther from shore show salinities and phase boundary temperatures more characteristic of normal sea water.

A downward salt flux exists at the bottom of the thawed layer, and the electrical conductivity of the interstitial water at one site shows evidence for a thin boundary layer there, within which the salt transport regime seems to change from convective to diffusive. Outside this layer, the salinity gradients are low, as would be expected in a well-developed convective regime.

Order of magnitude estimates of a characteristic interstitial water speed made from salinity and temperature data differ somewhat, but a few tenths of a metre per year seems reasonable. Although this is enough to make the transport of salt strongly convective (large salt Peclet number), the transport of heat is dominantly diffusive or "conductive" (small heat Peclet number). This is possible because the diffusivities of heat and salt differ by several orders of magnitude. Although no appreciable heat seems to be transported by convecting water, there is an extremely

important indirect effect in that the convective process seems largely to determine the phase boundary salinity and, therefore, temperature. This, in turn, determines the temperature gradient across the thawed layer and, therefore, the thaw rate (Harrison and Osterkamp 1978).

### Acknowledgements

The authors are grateful for the contributions of Masayuki Inoue and Robert Fisk and to many other people who helped in the field, usually under difficult conditions. The research was supported by the National Science Foundation under grants DPP 76-18399 and 77-28451 and by the NOAA-BLM Outer Continental Shelf Environmental Assessment Program.

### References

- CANNON, P.J. 1979. The environmental geology and geomorphology of the barrier island-lagoon system along the Beaufort Sea coastal plane. *In: Environ. Assess. Alaskan Cont. Shelf, Ann. Rep.*, vol. X, pp. 209-248.
- DOHERTY, B.T. AND KESTER, D.R. 1974. Freezing point of seawater. *J. Marine Res.*, vol. 32, no. 2, pp. 285-300.
- GOLD, L.W. AND LACHENBRUCH, A.H. 1973. Thermal conditions in permafrost — A review of the North American literature. *In: Permafrost — The North Amer. Contrib. to 2nd Int. Conf.*, Natl. Acad. Sci., pp. 3-23.
- HARRISON, W.D. AND OSTERKAMP, T.E. 1978. Heat and mass transport processes in subsea permafrost 1. An analysis of molecular diffusion and its consequences. *J. Geophys. Res.*, vol. 83, no. C9, pp. 4707-4712.
- HARRISON, W.D. AND OSTERKAMP, T.E. 1981. Details of a probe method for interstitial soil water sampling and hydraulic conductivity and temperature measurement. *Geophys. Inst. Rep. UAG R-280*, Univ. Alaska, Fairbanks, Alaska, 99701.
- HOPKINS, D.M. AND HARTZ, R.W. 1978. Shoreline history of Chukchi and Beaufort Seas as an aid to predicting offshore permafrost conditions. *In: Environ. Assess. Alaskan Cont. Shelf, Ann. Rep.*, vol. 12, pp. 503-575.
- ISKANDAR, I.K., OSTERKAMP, T.E., AND HARRISON, W.D. 1978. Chemistry of interstitial water from the subsea permafrost, Prudhoe Bay, Alaska. *In: Proc. 3rd Int. Conf. Permafrost*, vol. 1, pp. 93-98. Natl. Res. Council. Can., Ottawa.
- LACHENBRUCH, A.H. AND MARSHALL, B.V. 1977. Sub-sea temperatures and a simple tentative model for offshore permafrost at Prudhoe Bay, Alaska. U.S. Geol. Surv. Open-File Report 77-395.
- LEWELLEN, R. 1977. A study of Beaufort Sea coastal erosion, northern Alaska, final reports, *In: Environ. Assess. Alaskan Cont. Shelf, Ann. Rep.*, vol. 15, pp. 491-528.
- OSTERKAMP, T.E. AND HARRISON, W.D. 1980. Subsea permafrost: Probing, thermal regime and data analysis. *In: Environ. Assess. Alaskan Cont. Shelf, Ann. Rep.*, vol. 4, pp. 497-677.
- OSTERKAMP, T.E. AND HARRISON, W.D. 1981. Methods and equipment for temperature measurements in subsea permafrost. *Geophys. Inst. Report UAG R-285*, Univ. Alaska, Fairbanks, Alaska, 99701.
- OSTERKAMP, T.E. AND HARRISON, W.D. 1982. Temperature measurements in subsea permafrost off the coast of Alaska. *In: Proc. 4th Can. Permafrost Conf. Calgary, Alberta, 1981, Natl. Res. Council. Ottawa*, pp. 238-248.
- OSTERKAMP, T.E. AND PAYNE, M.W. 1981. Estimates of permafrost thickness from well logs in northern Alaska. *Cold Regions Sci and Technol.*, vol. 5, pp. 13-27.
- PAGE, F.W. AND ISKANDAR, I.K. 1978. Geochemistry of subsea permafrost at Prudhoe Bay, Alaska. U.S. Army Cold Regions Res. and Eng. Lab. Special Rep. 78-14.
- SELLMANN, P.V. 1980. Regional distribution and characteristics of bottom sediments of Arctic coastal waters of Alaska — Review of current literature. U.S. Army Cold Regions Res. and Eng. Lab. Special Rep. 80-15.
- UNESCO 1966. International oceanographic tables. UNESCO, Place de Fontenoy, Paris 7e, France; or Natl. Inst. Oceanogr. Great Britain, Wormley, Godalming, Surrey, England.