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SUBSEA PERMAFROST: PROBING, THERMAL REGIME AND DATA ANALYSIS

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I. SUMMARY OF OBJECTIVES, CONCLUSIONS AND IMPLICATIONS WITH RESPECT TO OCS DEVELOPMENT

The objectives of this study are to determine the occurrence, distribution and properties of subsea permafrost in Alaskan waters in cooperation with other OCSEAP investigators. Besides direct measurements, our program includes an effort to understand the basic physical processes responsible for the subsea permafrost regime as a basis for predictive models.

No field work has been done during this past year. Our major effort has been to continue data reduction and analysis and to begin writing papers, on subsea permafrost, for publication. The abstracts of these papers are included in the appendices.

This research has shown that it is likely that problems posed by subsea permafrost for offshore hot oil production will be greater than for permafrost problems onshore at Prudhoe Bay, because subsea permafrost is warmer, saltier and more easily disturbed, and because it is often associated with fine-grained soils.

II. INTRODUCTION

This work is part of the OCSEAP study of the distribution and properties of permafrost beneath the seas adjacent to Alaska and of processes that control the evolution of the subsea permafrost. The study involves coordination of the efforts of a number of investigators (RU 204, 271, 253, 255, 256, 473, 103, 407) and synthesis of the results of both field and laboratory work. Related work that is more focussed on the scientific problems of heat and mass transfer in subsea permafrost is primarily funded by the National Science Foundation and has been supported by OCSEAP logistics.

More information on specific objectives, and relevance to problems of petroleum development, is given in our previous annual reports.

III. CURRENT STATE OF KNOWLEDGE

The existence of subsea permafrost and some of its characteristics have been established at Prudhoe Bay, Barrow and other locations in the Beaufort, Chukchi and Bering Seas by drilling, probing, and seismic methods, as well as by studies of shoreline history, sea bed temperature, and regional geology. For example we know that ice-bearing permafrost occurs over most of the Beaufort Sea shelf, always at very shallow depths in near-shore areas. A hole 4 1/2 miles north of Reindeer Island, where it occurs only 8 m below the sea bed, suggests that it may be found close to the sea bed anywhere in the Beaufort Sea. Preliminary analysis of our field data suggests that, while gravels overlain by a thin cap of fine-grained sediment are common near Prudhoe Bay and to the west, the eastern portion of the Beaufort Sea appears to have more fine-grained sediments, with the gravels absent or overlain by a thicker cap of fine-grained sediments or overconsolidated clays. This zone of coarse-grained sediments extends roughly from the mouth of the Saganavirktok River seaward through Reindeer Island on the east to Oliktok Point on the west.

Theoretical concepts of the nature of the heat and salt transport mechanisms have begun to shed some additional light on the distribution and nature of subsea permafrost. For example, our diffusive heat and salt transport model (Harrison and Osterkamp, 1976) gave reasonable results when used to calculate the depth to ice-bonded permafrost in the consolidated clays north of Reindeer Island and our work on diffusive and convective salt transport mechanisms (Harrison and Osterkamp, 1978) has been used to develop a general hypothesis about subsea permafrost distribution (Hopkins and Hartz, 1978).

A few boreholes in the western Beaufort Sea have shown the existence of ice-bonded permafrost. The soils near Lonely are all fine-grained with the ice-bonded permafrost relatively close to the sea bed. It is known that the sediments are coarse to the east of Oliktok Point. The transition area from coarse to fine-grained sediments has not been *very* well" defined, nor has the effect of the Colville River on subsea permafrost conditions in Harrison Bay been determined.

"Several boreholes in the Norton Sound area suggest that subsea permafrost may be absent except possibly for some very favorable near-shore sites.

Drilling and probing data in the Chukchi Sea are very sparse. We have drilled several holes in the northern Chukchi Sea, near the Naval Arctic Research Laboratory (NARL), two holes north of Wainwright, one hole at Rabbit Creek and two holes near Kotzebue (Osterkamp and Harrison, 1982) .

During our Spring, 1981, field season we attempted to drill six holes in the general area between Cape Lisbourne and Point Lay. In all six holes, we encountered what appeared to be bedrock within 1/2 m of the sea bed, overlain by a thin layer of fine-grained sediment. This rock could not be penetrated with the equipment at hand, although a small modification of our present equipment would allow us to make boreholes in rock.

We attempted to drill one hole at Ogotoruk Creek near Cape Thompson at a site ≈ 75 m offshore in 6.40 m of water. The sea bed at this site was hard, apparently rock, but was not covered by sediment. The above data and our previous field data, while very sparse, suggest that there is rock at or near the sea bed in the area along the Chukchi Sea coast between Peard Bay and Cape Thompson. Rock was not found at NARL (NE of Peard Bay) nor at Rabbit Creek, SE of Ogotoruk Creek.

The presence of rock at or near the sea bed poses several problems for OCS development. First, the rock cannot. be assumed to be a good foundation material for structures unless it can be shown that it does not contain segregated ice. Second, laying pipelines in this rock could be extremely difficult. Third, if the rock is as widespread as our sparse data suggests, it may be very difficult to obtain local gravel for construction of docks, causeways, artificial islands, etc.

The general approach of these studies has been to make detailed measurements in key representative areas while using the theoretical concepts to make wider inferences on the basis of regional geology, shoreline history, and heat and salt transport models. This approach has been used because the thousands of kilometers of Alaskan coastline potentially subject to subsea permafrost conditions cannot be studied in detail. As noted earlier, coordinated efforts among a number of investigators using different techniques have been made. Our past program has been to investigate subsea permafrost conditions with the help of lightweight probes and sampling apparatuses, and to infer larger-scale conditions from other data, and from theory.

The portable nature of the equipment that was developed has allowed us to drill boreholes in Norton Sound, Kotzebue Sound, Hope Basin, northern Chukchi Sea, Elson Lagoon, Lonely, Harrison Bay, Prudhoe Bay and at other offshore sites. A number of boreholes have also been made in the barrier islands off the Beaufort Sea. By concentrating our efforts at a few sites in these areas, we hope to be able to extrapolate the information obtained over much of the areas.

Many questions remain to be answered and our measurement, sampling, and theoretical programs coupled with the results of other OCSEAP investigators are capable of providing substantial answers to these questions.

IV. STUDY AREA

No field work was carried out during this past year. Our efforts since 1975 have included studies in the Bering, Chukchi and Beaufort Seas. We are presently concerned with data reduction, analysis and publication of our results.

V. METHODS AND RATIONALE OF DATA COLLECTION

Although there have been some refinements, our methods have not changed greatly from those described in our report (RU 253, Osterkamp and Harrison, 1980). We have recently published two papers (Harrison and Osterkamp, 1982; Osterkamp and Harrison, 1982) that give a fairly complete description of most of the methods. Successful measurements of the pore water pressure profile in the sediments were conducted during 1981 at Prudhoe Bay for the first time and are included in last year's annual report. The results are discussed further in a paper submitted to the

Fourth International Conference on Permarost (July, 1983 in Fairbanks] and a copy is given in Appendix A.

VI, VII, VIII. SUMMARY OF RESULTS AND DISCUSSION

Subsea permafrost (SSP) exists in areas that were once emergent and exposed to arctic air temperatures similar to today's. Subsequent submersion, at least in Alaska, has usually been by a combination of eustatic changes in sea level and by shoreline erosion, which subjects the permafrost to a relatively warm and salty surface boundary condition. Because of this complex history, efforts to formulate predictive SS? models are still in their infancy, partly because of the formidable difficulties caused by its essentially transient nature, and the important and complex role of salt. To develop predictive models one must know the past as well as present boundary conditions, the relevant material properties, and the heat and mass transport processes that control the rate of evolution. This requires a knowledge of the lithology. It is, therefore, not surprising that most of what is known about SSP on the shelves of Alaska is from rather direct interpretation of still relatively unsynthesized data.

These data came from several sources: drilling and sampling at depths up to 100 m (Lewellan, 1975; Sellmann, 1980); deep exploration holes with geophysical and temperature logs still largely proprietary (Osterkamp and Payne, 1981); reflection and refraction measurements carried out from small boats (Rogers and Morack, 1978); proprietary transient electromagnetic measurements carried out off the sea ice; interpretation of the first returns from deep seismic data (Sellmann and Neave, 1982), studies of shoreline position and geologic histories (Hopkins and Hartz, 1978), and reconnaissance drilling and probing measurements to depths up to 40 m (Osterkamp and Harrison, 1982; Harrison and Osterkamp 1982). The references given are but a sample of the data in the public domain. Not referenced but *extremely* important are the rapidly multiplying sea bed investigations by industry. Similar data are available from the Canadian side of the Beaufort Sea.

After submersion, a thawed layer develops beneath the sea bed, and geothermal heat thaws the permafrost from below. The properties and thickness of the sea bed thawed layer will play a key role in offshore foundation and pipeline design. To the extent that shoreline retreat rate is constant, distance offshore is proportional to time, so the evolution of the thawed layer over longer time periods can be studied by moving farther offshore. In some respects, the SSP in the nearshore region, which marks a transition from subaerial to subsea conditions, is the most interesting, and it has always been a focus of our research. There the SSP is very young, formed as a result of shoreline erosion rates from 1 to 10 m a^{-1} in much of Arctic Alaska. Its response to the new warm and salty surface conditions may be very rapid, particularly just beyond the 2 m isobath where the sea ice does not grow to the seabed causing a relatively low mean temperature.

The rate of thaw beneath the sea bed varies by orders of magnitude depending upon the soil type. Where impermeable sediments exist at the sea bed, the thaw rate is extremely slow, sometimes less than 10 m after 10,000 years, while greater thaw can take place in only 50 years in permeable sediments. This situation is obviously due to fact that in the Beaufort Sea of Alaska the mean annual sea bed temperature is negative, so salt must play a critical role in the thaw process. Some of the salt is present before submersion, but at least in some areas most of it enters from the sea water as thawing progresses (Harrison and Osterkamp, 1982). We showed some years ago how a key factor in the evolution of the thawed layer in sufficiently permeable sediments is convection of the pore water, which gives *rise* to efficient salt tansport and fast thawing (Harrison and Osterkamp, 1978).

Our role in Alaskan SSP research has consisted of two parts: (1) reconnaissance studies of SSP conditions in new areas using lightweight drilling and probing equipment, and (2) studies of heat and mass transport processes in SSP. The former program was funded by OCSEAP and a final report is in preparation. The latter program, is the only study so far that has focussed on the complex physics and chemistry of the transport processes. It is an argument in favor of basic research that our work on transport processes, which may at first have appeared to have only academic interest, has recently seen application in the design of gathering pipelines for offshore oil wells.

Although we carried out the first SSP investigations at Prudhoe Bay in 1975 using a drill rig with conventional soil sampling capability (Osterkamp and Harrison, 1976), we early felt the need for the development of lightweight jetting and probing equipment that could be inexpensively deployed. This development has been a significant component of our

NSF-funded research. Probing equipment that can be deployed by snow-machine, and can reach tens of meters to the bottom of the thawed layer beneath the sea has been used for most of the research. With it we can measure temperature, hydraulic conductivity and pressure profiles, obtain pore water samples through the thawed layer, and measure its thickness. Descriptions of the probing techniques, which are rather unconventional, are given by Harrison and Osterkamp (1981, 1982) and Osterkamp and Harrison (1982). They are actually better suited than a drill rig for measurement of temperature and pore water pressure.

Some of our results for rather fine-grained soils off Tekegakrok Point, Elson Lagoon, are shown in Figures 1 and 2 (Osterkamp and Harrison," 1978). The situation is not simple. Phase boundaries are ill-defined, possibly because a significant amount of salt was present before submersion. A surprise. is that the highly saline layer present near shore is missing farther offshore. Because the shoreline retreat rate here is roughly 2 m a^{-1}, the dissipation of the saline layer must have been relatively rapid, much more so than indicated by the low permeabilities we measured (which should be too low to permit convection). Molecular diffusion is too slow to account for the effect. A highly permeable aquifer that we found during the observations may provide a clue; if these aquifers are common, it is incorrect to think in terms of a continuum.

Our best data set comes from the relatively permeable soils of the Prudhoe Bay West Dock area, where we know that the phase boundary marking the bottom of the thawed layer is sharp (usually defined mechanically to within about 0.1 m), and has a curiously uniform shape, as outlined in Figure 3. The parabolic shape beyond 440 m from shore can be understood in terms of a constant shoreline retreat rate (of about 1 m/a) and a $(\text{time})^{1/2}$ thaw rate dependence, and the lack of significant thawing inshore of 400 m in terms of low mean annual seabed temperature caused by freezing of sea ice to the sea bed. We do not yet understand the rapid thaw rate between 400 and 440 m. No plausible sea bed temperature model can by itself account for this effect.

Another surprise is the simplicity of the phase boundary. It seems to be much smoother, and to have an astonishingly more uniform temperature (close to -2.40°C) than we would expect given inhomogenities in soil ice content, in initial thermal conditions (because of lakes), and especially in sea bed temperature and salinity conditions.

A sample West Dock temperature profile through the thawed layer is shown in Figure 4. These profiles tend to be quite linear, except for seasonal effects near the sea bed. The linearity indicates a small Peclet number for heat transport and therefore puts an upper limit of a few tenths of a meter per year on the pore water convective velocity (Harrison and Osterkamp, 1978; Osterkamp and Harrison, 1982). Small observed non-linearities may in fact be due to convection, but the effect is too small to be completely convincing. We"have measured the thermal conductivity profile through the thawed layer by a borehole heating technique (Osterkamp and Harrison, 1980). Sample salinity profiles are shown in Figure 5. They tend to show small gradients, which imply well-developed convection. Of special interest is the small but characteristic salinity decrease of a few percent at the bottom of the thawed layer. We interpret it as a thin diffusive boundary layer in a strongly convecting system (Harrison and Osterkamp, 1982). We expected it on theoretical grounds, and its observation is probably one of our major field achievements. During our last field operation we were finally successful in measuring tiny non-hydrostatic pore water pressure gradients (in one hole), after several years of frustrating attempts (Osterkamp and Harrison, 1982) (Figure 6). This success was not anticipated at the beginning of our program, and is the most direct evidence so far for pore water convection. The observations have uncovered several other features, such as evidence for rather rapid seasonal changes in pore water salinity near the seabed. This Indicates the desirability of year-round measurements.

Although there seems to be little doubt that pore water motion occurs, the details of the motion are not known, nor, for certain, the driving mechanism. We showed (Harrison and Osterkamp, 1978) how gravity could be the mechanism, ultimately because relatively fresh and therefore buoyant water is generated by thawing at the base of the thawed layer. (Density is completely controlled by salinity at the existing temperatures). The most obvious alternative to gravity is surface waves, which induce pore water motion. It has been shown that the dominant resulting transport mechanism is the mechanical dispersion familiar to ground water hydrologists, and the appropriate theory has been worked out (Harrison and others, 1983). This work is in press, and an abstract is in Appendix B. The process is probably unimportant at Prudhoe Bay, because the permeability

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is too low, but it is probably important in chemical and biological processes elsewhere *where* permeability is higher, such as the eastern U.S. shelf.

Additional possible driving mechanisms have been considered, but gravity still remains the most likely, and most of our analysis has been for that case (Harrison, 1982). Two interpretative approaches have been followed. One is semi-quantitate and uses several basic relationships, such as Darcy's law and the conservation of energy, to reconcile measured or estimated values of thaw rate, pressure gradient, hydraulic conductivity to porosity ratio, and the small salinity jump across the boundary layer. This approach leads to the following conclusion: Given the uncertainties in the data, gravity driving seems quite reasonable, but if so, the measured values of hydraulic conductivity and of the salinity jump are too low. The results of these analyses have been prepared for presentation at the Fourth International Conference on Permafrost (Appendix A).

The second approach is to carry out a complete numerical calculation of the pore water velocity field. This was originally part of the proposal for our current NSF research, but it is a major undertaking and for funding reasons had to be deleted. It was subsequently funded separately in a proposal to NSF in which Professor Swift was the P*I* (Swift and others, 1980). The formulation of the theory has been part of our NSF project, and is summarized in Appendix C, most of which is taken from Harrison (19821. A manuscript giving the first numerical results has been prepared. An abstract is in Appendix D. There are two main difficulties; first, the Rayleigh number is extremely high because of the small molecular diffusivity for salt, and second, the boundary conditions are extremely complex, the lower boundary being in motion, and the salinity and temperature there being coupled by phase equilibrium. Despite the simplifications made in these first calculations, we think that they have shown the morphology of the velocity field correctly. The upward motion is limited to narrow plumes. This is important to the planning and interpretation of the proposed field observations. For example, it suggests high probability that measured pressure profiles will usually indicate downward motion, as does the existing one in Figure 6.

Thermal Models

One crude but very useful approach to a certain class of SSP problems is to ignore the details of the mass transport process, and to assume that heat transport is primarily conductive (as field observations seem to indicate). Then heat conduction theory can be used to describe the permafrost response to changing boundary conditions, such as those induced by a changing shoreline position. Some of the types of calculations that can be carried out are given in Osterkamp and Harrison (1982). Obviously there are some serious limits. For example, one cannot predict the rate of development of the thawed layer from a sea bed boundary condition this way, but at least in the Prudhoe Bay West Oock area the difficulty can be circumvented by using the measured phase boundary temperature. A model giving the time dependence of the thawed layer thickness and the depth of the base of the permafrost can then be solved. This has been done by Lachenbruch and others (1982) for Prudhoe Bay, and by us (Harrison and Osterkamp, 1977) for the Chukchi Sea. In a case such as the latter, where thermal properties on boundary or initial conditions are unknown, one uses extreme values to bracket the possible range of behavior.

We have acquired temperature data from boreholes in the Bering, Chukchi and Beaufort Seas which can be used for thermal models. Additional data have been acquired by several programs besides ours, and we have collected most of them on a computer file. The data fall into two categories, near shore where freezing of sea ice to the sea bed and the two dimensions"l effect of the nearby cold land are important, and farther offshore where these effects are negligible. The near-shore theory is obviously more difficult, and some of it, a generalization of that of Lachenbruch (1957), has already been worked out by us but not yet published.

All of the thermal offshore calculations done so far (including our own) contain errors because they neglect the effect of the thermal inertia of the ground beneath the permafrost. Another difficulty is that all present calculations, with the exception of our unpublished near shore ones, do not take into account the presence of salt in ice-bearing permafrost, which is known to be present from several lines of evidence. By distributing the melting point over a finite temperature range, even a small amount of salt can greatly increase the effective bulk heat capacity, and lead to *a* quite different temperature response than the salt free case. The abstract of a paper describing some of the results of thermal calculations in subsea permafrost is given in Appendix E.

IX. ACKNOWLEDGEMENTS

We wish to acknowledge the assistance of Robert Fisk and Victor Gruel in this research including much work under difficult to impossible field situations, the NOAA helicopter crews and the Geophysical Institute machine shop personnel. Logistical support was provided by OCSEAP. The National Science Foundation supported parts of this research under grant DPP-77-28451.

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Figure 1. Temperature profiles off Tekegakrok Point Elson Lagoon. In this and following figures, hole numbers are distances from shore *in* meters.

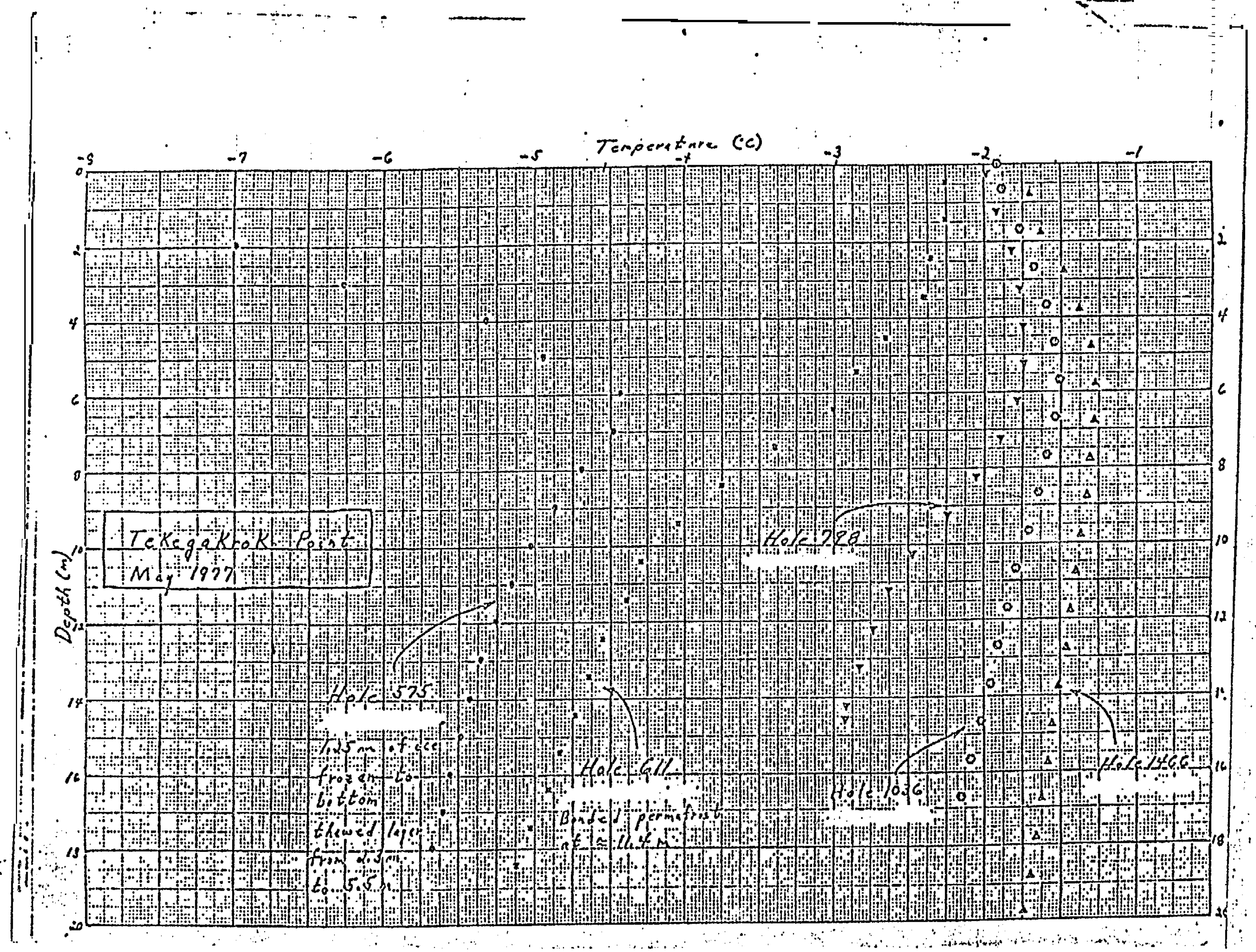
Figure 2. Pore water salinity profiles, expressed in terms of freezing point depression, off Tekegakrok Point, Elson Lagoon.

Figure 3. Configuration of the base of the thawed layer at West Dock, Prudhoe Bay. Seaward of 440 m, the thawed layer thickness (Y) closely follows $Y = 1.147\sqrt{x - 6}$, where x is distance from shore, expressed in meters.

Figure 4. Sample West Dock temperature profile through the thawed layer.

Figure 5. Sample West Dock pore water electrical conductivity profiles through the thawed layer.

Figure 6. West Dock pore water pressure profile.



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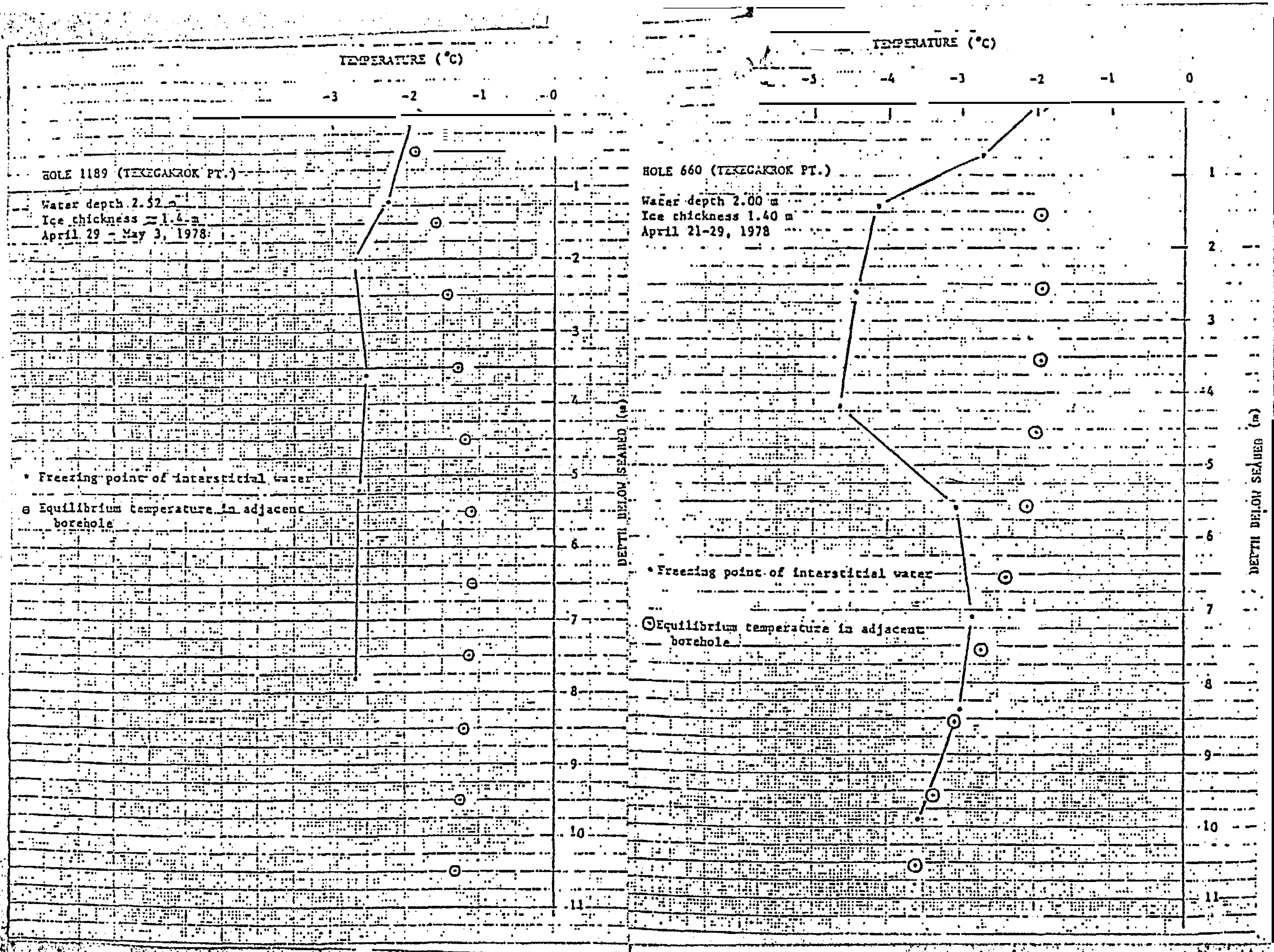


Figure 2. Pore water salinity profiles, expressed in terms of freezing point depression, off Tekegakrok Poi[illegible] [illegible]lson Lagoon.

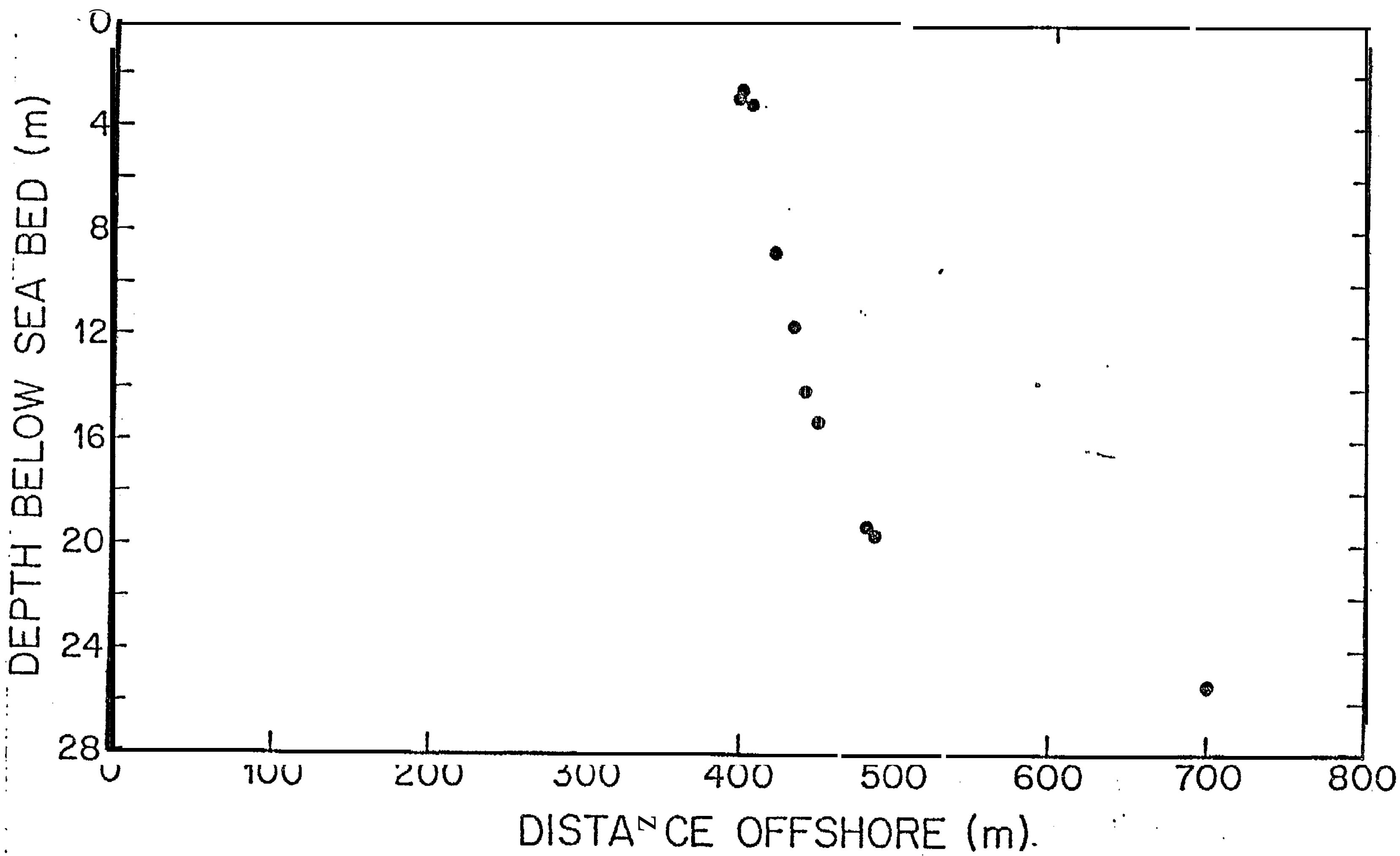


Figure 3. Configuration of the base of the thawed layer at West Dock, Prudhoe Bay. Seaward of 440 m, the thawed layer thickness (Y) closely follows $Y = 1.147 \sqrt{x - 276}$, where x is distance from shore, expressed in meters.

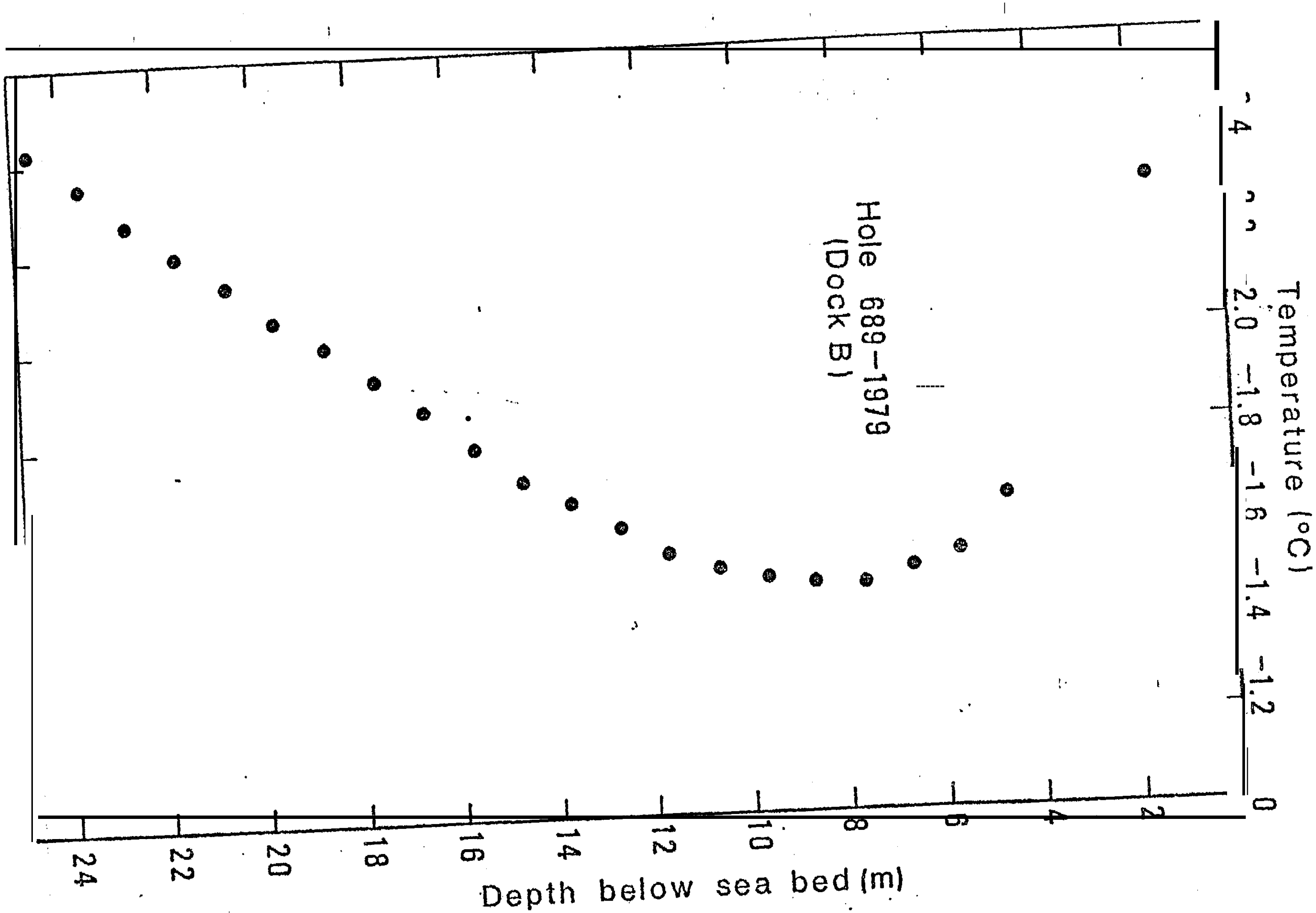
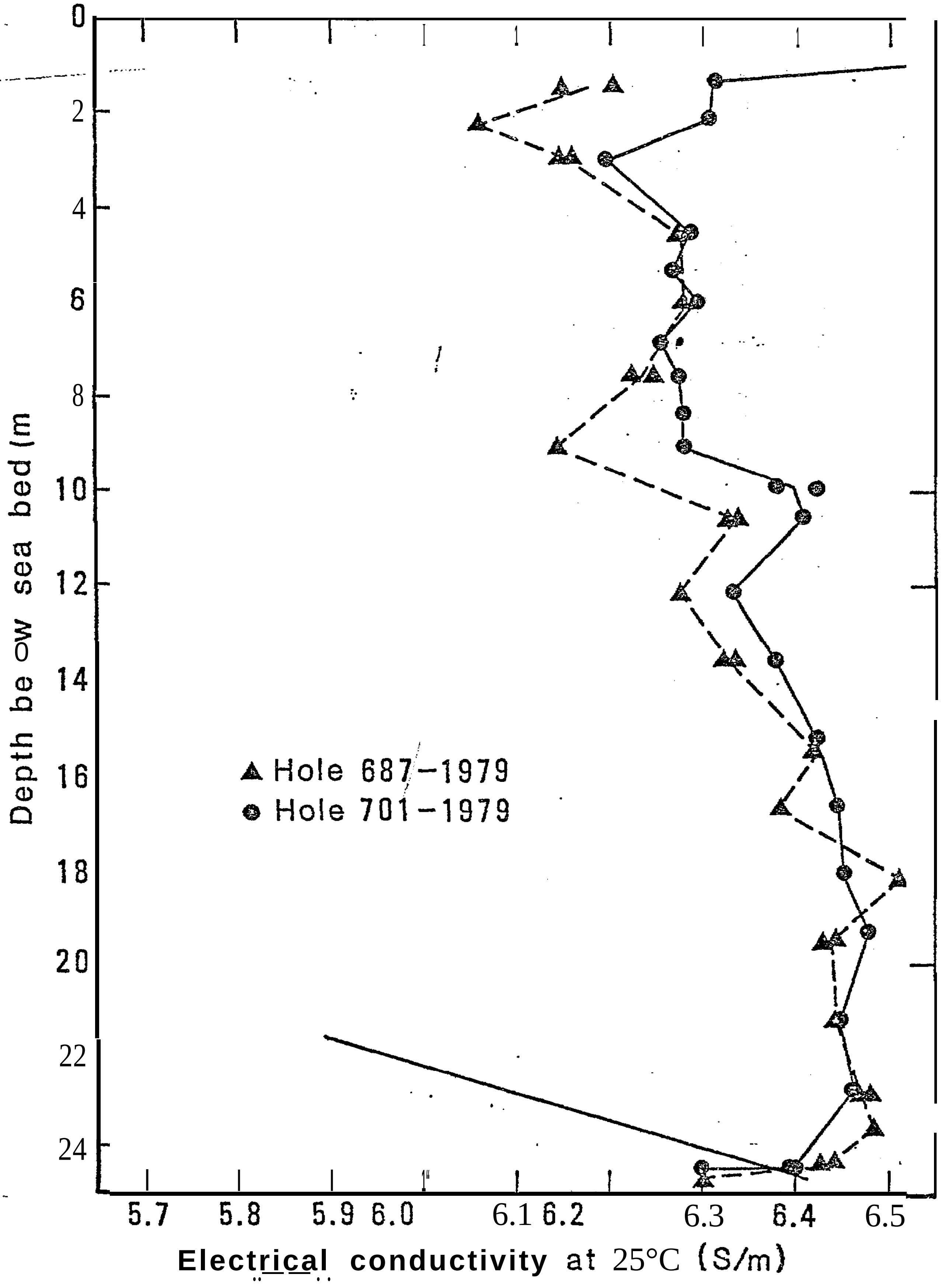


Figure 4. Sample West Dock temperature profile through the thawed layer.



F gure 5. Sample West Dock pore water electrical conductivity profiles through the thawed layer.

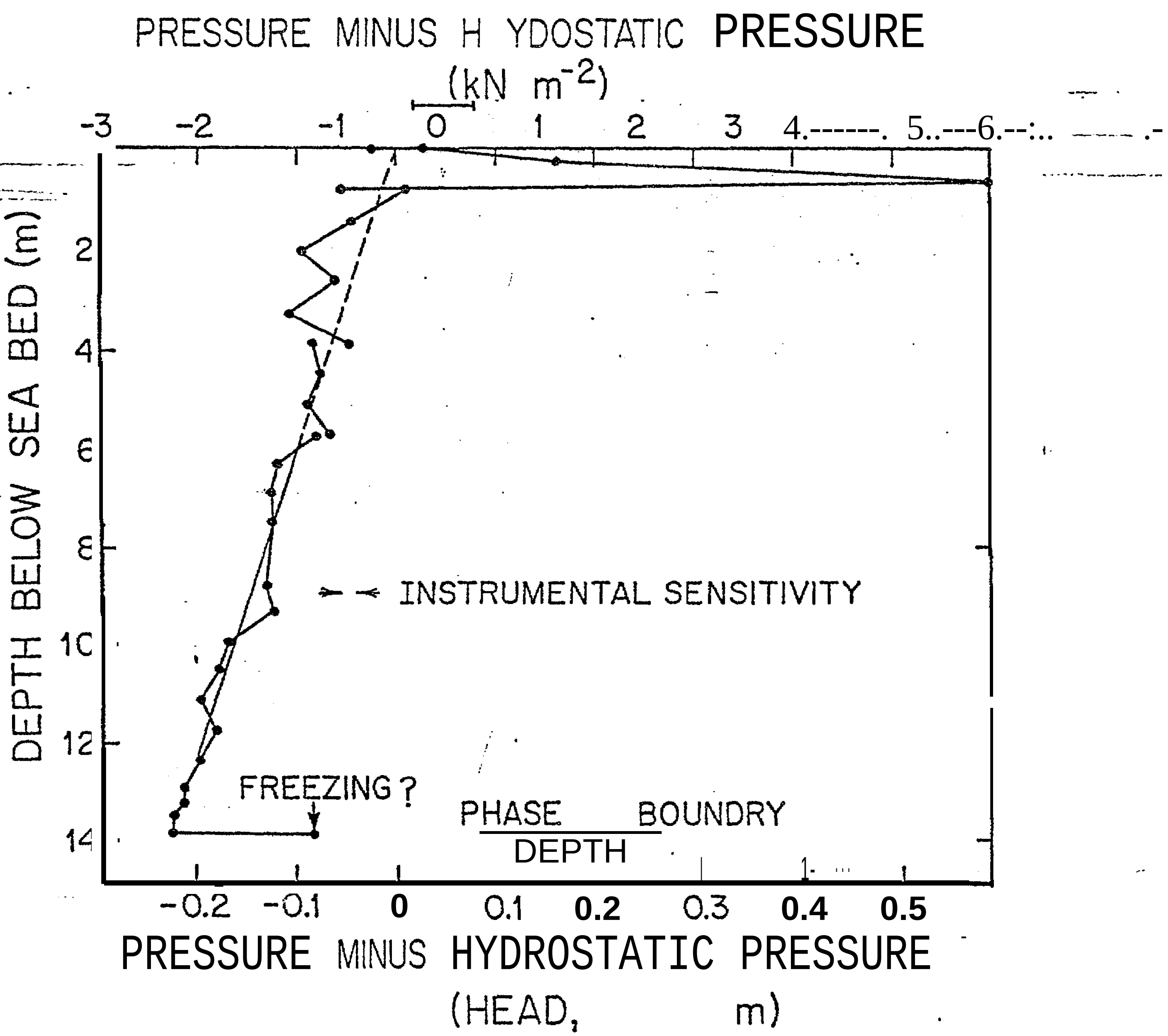


Figure 6. West Dock pore water pressure profile.

XI. APPENDICES A THROUGH E

APPENDIX A

{Submitted for presentation at Fourth International Conference on "
Permafrost, Fairbanks, Alaska, July 1983)

4081

HEAT AND SALT TRANSPORT PROCESSES IN THAWING SUBSEA PERMAFROST AT PRUDHOE BAY, ALASKA

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In the negative subsea temperature regime characteristic of Prudhoe Bay, Alaska, the rate of development of the thawed layer underlying the sea bed is largely controlled by salt transport processes through it. The heat and mass transport processes that act in a given *region* depend on lithology; for an example, it is known that thawing is rapid in relatively permeable sediments, but extremely slow where an impermeable cap of over-consolidated clay exists at the sea bed. The present state of our understanding of the transport processes operating in a relatively permeable *region*, near the Prudhoe Bay West Dock, is discussed in terms of field observations and theory. The observations consist of shore line . erosion rates, thickness of the thawed layer as a function of distance from shore, lithology, and vertical profiles of temperature (discussed in a companion paper), pore water salinity, hydraulic conductivity, and pore water pressure. Most of the profile data were obtained by a probe technique developed for this research. The pore pressure data seem to be the first of this kind available.

The salinity profiles indicate a well-developed *convective* salt "transport regime [Peclet number for salt >>1) in the thawed layer, and the

known thaw rate puts lower limits on the characteristic pore water velocity. The temperature profiles tend to be linear (Peclet number for temperature <<1); this, and the pore water pressure profile, set upper limits on the velocity, which seems to be of order 0.1 to 1 meter per year. Observations of a thin salt-diffusive boundary layer at the bottom of the thawed region gives further information about the velocity.

Theory shows that the necessary conditions for-gravitational instability of the pore water to convective motion exist, the driving force being the relatively fresh, buoyant water generated at the bottom of the layer by thawing. The matter of the driving mechanism is not yet settled however, because the frictional energy dissipation estimated from Darcy's law exceeds the available gravitational potential energy. This suggests that another driving mechanism may exist or that estimation of frictional energy is too high because the hydraulic conductivity measurements have given values systematically too low. One other possible driving mechanism, the forcing of the pure water motion by surface wave action, has been investigated theoretically and found to be unimportant under Prudhoe Bay conditions.

APPENDIX B

4082

(In press, Journal of Geophysical Research)

4710 (Chemical Oceanography)

A WAVE INDUCED TRANSPORT PROCESS IN MARINE SEDIMENTS

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We show how surface wave action can increase the rate of transport of solutes into a sandy sea bed by orders of magnitude via a mechanism known as mechanical dispersion. It is most effective for large sediment permeability and thickness, high surface wave amplitude, and shallow water. A method for setting up the appropriate transport equation is given; it is valid when dispersion is well developed. The dispersion term contains two mechanical dispersion parameters that can be estimated roughly from existing data when the sediments are well sorted. The dispersion can be inhomogeneous and anisotropic even in homogeneous; isotropic sediments. It may be an important transport mechanism in sediments on the eastern U.S. shelf, but not in tune thawing of subsea permafrost beneath Prudhoe Bay, Alaska. (Transport, sediment, waves).

4705, 4770

APPENDIX C

FORMULATION OF A MODEL FOR HEAT AND SALT TRANSPORT THROUGH THE THAWED LAYER

This is a description of the formulation of a theory of heat and mass transport through the thawed layer, appropriate when Darcy's law is obeyed. It provides the basis of the numerical calculations referred to in the text. We begin with a definition of symbols - unfortunately numerous.

Coordinates:

- x horizontal, perpendicular to shore
- y vertical, positive clown, zero at sea bed
- z horizontal, parallel to shore
- t time
- $\hat{e}_g$ unit vector along y-axis
- Y phase boundary y-coordinate

Heat:

- Pe_T Peclet number for heat transport
- T temperature (^oC)
- κ thermal diffusivity
- γ porosity times ratio of water and bulk heat capacities
- h latent heat per unit volume of the porous medium
- K_1 thermal conductivity above phase boundary
- K_2 thermal conductivity below phase boundary
- $T_0 = T|_0$ sea bed temperature
- $T_Y = T|_Y$ phase boundary temperature
- G equilibrium geothermal gradient in permafrost onshore
- T_s ground surface temperature onshore
- T_f freezing temperature of *overlying* sea water with salinity S_0.

Salt:

- Pe_S Peclet number for salt transport
- S porewater salinity
- S_r a reference salinity, zero of salinity scale
- κ_S diffusivity of salt
- S' $S - S_r$
- $S_0 = S|_0$ sea bed salinity
- $S_Y = S|_Y$ phase boundary salinity
- S a characteristic salinity used in defining a Rayleigh number
- ΔS salinity change across the boundary layer at the phase boundary

Mass flow:

$\underline{v}$	actual water velocity
$v_{x,y,z}$	velocity components
ϕ	"porosity
k	hydraulic conductivity ÷ ϕ
p	pressure, expressed as a head
ρ	density
ρ_r	a reference density corresponding to the reference salinity S_r
a	a constant expressing the dependence of ρ on S
f (T)	a function of T expressing the dependence of ρ on T
P'	P - y
v_c	a characteristic water velocity
v_T	the flux velocity
Ra	Rayleigh number
v	velocity along the phase boundary
b	boundary layer thickness

The theory is summarized in a fairly general way, so the subsequent approximations will be more obvious. It is as follows:

$$\nabla\cdot(\phi \underline{v}) = 0, \qquad \text{Incompressibility,}$$

$$\underline{v} = -k(\nabla P - \rho/\rho_r\, \hat{e}_g), \qquad \text{Darcy's law,}$$

$$\nabla\cdot(\kappa_s \nabla S) - \underline{v}\cdot\nabla S - \frac{\partial S}{\partial t} = 0, \qquad \text{Salt transport,}$$

$$\kappa \nabla^2 T - \gamma\, \underline{v}\cdot\nabla T - \frac{\partial T}{\partial t} = 0, \qquad \text{Heat transport,}$$

$$\rho/\rho_r = 1 + a(S - S_r) + f(T), \qquad \text{Equation of state.}$$

There are two main assumptions in these equations, that the sediments are isotropic, and that no inertial terms are needed in Darcy's law. The latter needs to be checked after more is known about the velocity solution. Additional assumptions are as follows: that the porosity (ϕ) and salt diffusivity (κ_s) are constant, that heat transport is primarily conductive

($\gamma\, \underset{\sim}{v}\cdot\nabla T$ negligible) , and that the effect of temperature variations on density is negligible ($f(T)$ negligible) . The latter two assumptions seem to be quite valid. Further simplification results from combination of the second and fifth equations. We then have

$$\nabla\cdot\underset{\sim}{v} = 0, \tag{1}$$

$$v = -k(\nabla P' - a\, S'\, \hat{e}_g), \tag{2}$$

$$\kappa_S \nabla^2 S' - \underset{\sim}{v}\cdot\nabla S' - \frac{\partial S'}{\partial t} = 0, \tag{3}$$

$$\kappa \nabla^2 T - \frac{\partial T}{\partial t} = 0, \tag{4}$$

in which the unknowns are $\underset{\sim}{v}$, P' , S' and T. In solving these equations the hydraulic conductivity to porosity ratio (k) would "be assumed constant as well. The most important aspect of these simplified equations is that the temperature enters only into the last, and is therefore decoupled from the other variables as far as the governing equations are concerned. However, we will see that it is still coupled via the boundary conditions.

Boundary conditions apply at the sea bed and at the phase boundary. In the most general case they involve derivatives normal to these surfaces. Except close to shore it is probably a reasonable approximation to assume these surfaces are horizontal, thereby simplifying the complex geometry to a horizontal infinite layer. z-dependence of the boundary conditions is also ignored; this is also a reasonable approximation. At the phase boundary the boundary conditions then become

$$v_Y\big|_Y = 0, \qquad \text{Impermeable boundary,} \tag{5}$$

$$S\big|_Y \propto -T_{|Y}, \qquad \text{Phase equilibrium,} \tag{6}$$

$$S\big|_Y \frac{dY}{dt} = -\kappa_S \frac{\partial S}{\partial y}\bigg|_Y, \qquad \text{Stefan condition for salt,} \tag{7}$$

$$h\frac{dY}{dt} = -K_1 \frac{\partial T(y<Y)}{\partial y}\bigg|_Y + K_2 \frac{\partial T(y>Y)}{\partial y}\bigg|_Y, \qquad \text{Stefan condition for heat.} \tag{8}$$

The third condition implies that there is no salt transport below the phase boundary. The second sometimes needs to be corrected for the effect of pressure. At the sea bed the boundary conditions are

$$v_x\big|_0 = 0, \tag{9}$$

$$S\big|_0 = \text{prescribed function of } x \text{ and } t \tag{10}$$

$$T\big|_0 = \text{prescribed function of } x \text{ and } t.$$

The first of these implies a permeable sea bed, where the overlying free water does not sustain any pressure gradient. It can be stated in the equivalent form $P'\big|_0 =$ constant. In fact, travelling waves on the surface of the ocean do cause sea bed pressure variations that induce pore water motion and contribute to salt transport. The phenomenon has been investigated theoretically by Harrison and others (1982) for the case in which gravity-driven convection is absent, and found to be negligible when the hydraulic conductivity has the magnitude characteristic of Prudhoe Bay. It therefore seems likely that gravity driven convection is the dominant process, so the constant pressure boundary condition should suffice.

The problem is now to simplify the theory by eliminating temperature in as realistic a fashion as possible. First, because in the simplest approach the phase boundary motion is neglected, it is convenient to eliminate $\frac{dY}{dt}$ by combining equations (7) and (8):

$$-\frac{\kappa_s}{S_Y}\frac{\partial S}{\partial y}\bigg|_Y = -\frac{K_1}{h}\frac{\partial T(y<Y)}{\partial y}\bigg|_Y + \frac{K_2}{h}\frac{\partial T(y>Y)}{\partial y}\bigg|_Y \qquad (11)$$

Beyond the ramp area, 440 m from shore, the gradients in equation (11) could be calculated in closed form from Stefan Neumann theory, because it is known from field evidence that To and T_Y are fairly constant. The unknown temperature Ty could then be eliminated by the use of Equation (6). (T_Y is known from field observations, but it is an objective of the theory to calculate it correctly.) The problem would then be reduced to solving equations (1), (2) and (3} subject to boundary conditions given by equations (9) and (10), which apply at the sea bed, and by equations (5) and (11), which apply at the phase boundary (y = Y}.

In the numerical calculations performed so far, a simpler approach has been followed. The second term on the right side of equation {11) has been neglected, and the first replaced by $-\frac{K_1}{h}\left(\frac{T_{Y_1} - T_0}{Y}\right)$; this is valid in the limit of very large h and is roughly valid at the Prudhoe Bay West Dock. Rearrangement, and the use of equation (6) then give

$$S_Y\left(-1 + \frac{S_Y}{S_f}\right) + \left[\frac{\kappa_s h Y}{K_1(-T_0)}\right]\frac{\partial S}{\partial y}\bigg|_Y = 0$$

rather than equation (11), for the salinity boundary condition at y = Y. It contains two parameters, the salinity (S_f) of water that would begin to freeze at the sea bed temperature To, and the combined constants in the square brackets. S_Y is to be determined as part of the solution of the problem; T_Y can then be found from equation (6). The thaw rate is finally found from equation (7).

In the numerical calculations performed so far, Y has been treated as a constant, both *in* time and space, although in fact it should be calculated as a function of x for each time step. This problem is probably not too severe because the calculated pore water velocity greatly exceeds $\frac{dY}{dt}$, and because Y is observed in the field to be smoothly varying.

APPENDIX D

(Abstract of manuscript for journal publication]

Convective Transport of Brine and Thaw of Subsea Permafrost: Results of Numerical Simulations

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Abstract

A two-dimensional numerical simulation is used to investigate the circulation of brine in the porous bottom sediments. on the continental shelf of the Arctic Ocean. The brine motion is assumed to be governed by D'Arcy's Law and is effective in transporting salt to the ice rich sediments underlying the thaw region. The results of the simulation show that the flow is characterized by a few very narrow plumes consisting of upward moving relatively fresh water released from melting of fresh water "ice. The region between the plumes 1s characterized broad downward moving flow. The salinity density is nearly uniform, except within the plumes and along the very thin bottom boundary layer. The code indicates a strong tendency for the upward plumes to coalesce so the number of upward plumes decreases with time.