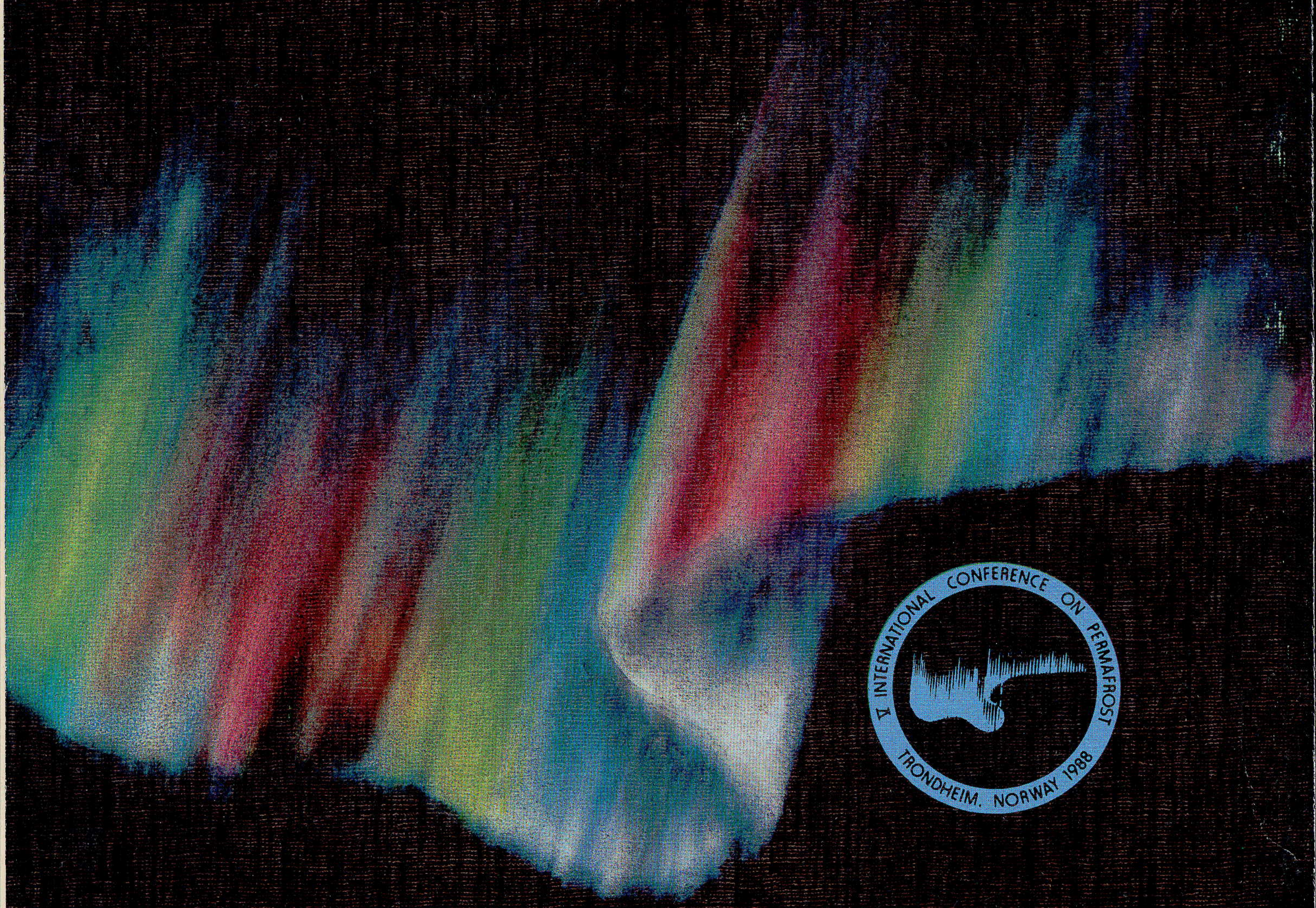
V International Conference on Permafrost
in Trondheim, Norway, August 1988



V INTERNATIONAL CONFERENCE ON PERMAFROST
TRONDHEIM, NORWAY 1988

PERMAFROST

VOL.3

TAPIR

V INTERNATIONAL CONFERENCE ON PERMAFROST
TRONDHEIM, NORWAY 1988

PERMAFROST

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PREFACE – VOLUME 3.

This volume, along with Volume 1 and 2 which contain the technical papers presented at the Conference, makes up the entire set of the Fifth International Permafrost Conference Proceedings.

In accordance with the information given in the IPA Council Meeting succeeding the Conference, this volume contains the Special Session Presentations, Additional Papers, Poster Presentations and their synopses, as well as a List of Participants.

The Norwegian Organizing Committee appreciate very much the cooperation and support provided by the Norwegian publication "Frost Action in Soils", from The Royal Norwegian Council for Scientific and Industrial Research. This third volume is issued as volume 27 of Frost Action in Soils.

The Organizing Committee will use the opportunity to thank all contributors to the proceedings presented. They cover many engineering and scientific disciplines and present research activities and engineering solutions from many countries.

The Conference comprised 289 papers, 8 Special Session Presentations and 37 Poster Presentations. 305 participants were registered.

The Organizing Committee is indebted to all involved as chairmen, speakers, discussion participants, technical tour guides and to the local Trondheim organizers for their successful efforts.

Finally, the cooperation and assistance given by Tapir Publishers are greatly appreciated.

Kaare Flaate | Kaare Senneset

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PERMAFROST TEMPERATURE AND THE CHANGING CLIMATE

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Synopsis

Temperature profiles in continuous permafrost to depths of a few hundred meters contain a faithful record of change in surface temperature during the past few centuries. When interpreted with heat-conduction theory, this little known source can provide important information on patterns of contemporary climate change. Precision measurements in oil wells in the Alaskan Arctic indicate a variable but widespread warming (typically 2°–4°C) at the permafrost surface during the 20th century. Recent statistical studies of 100 years of weather records from the North American Arctic suggest similar large changes have occurred in the air temperature although they might have started somewhat earlier. The permafrost warming is conspicuous and easily measured, and it is occurring at high latitudes where anthropogenic climatic change is expected to be greatest and first observable. Although permafrost is well-suited to the detection and monitoring of such changes, little is presently known about either the geographic extent or cause of its warming.

INTRODUCTION

We all know that permafrost is associated with a cold climate, but how well do we understand this association? What is the relation between thermal conditions in permafrost and climatological parameters? What can permafrost tell us about cold climates of the past? How will changing climate affect permafrost, and changing permafrost affect climate? And finally, what will be the effects of such changes on the interacting landscapes and biological systems?

Studies in natural and engineering sciences in permafrost regions have traditionally focused on one or two of three physically distinct domains: the atmosphere, the seasonally thawing active layer, or the permafrost. Each is governed by different physical processes and consequently each is usually studied by people with backgrounds in different disciplines. To answer the foregoing questions we must develop new understanding of the interactive processes that take place across the domain boundaries. However, in some cases our disciplines are still so separated that a useful beginning can be made by applying procedures that are well known on one side of the boundary to questions being asked about the other side. An example is the subject of this paper: the use of temperatures measured deep in permafrost to reconstruct the recent history of temperature change at the earth's surface. Although the general method has been familiar to solid-earth geophysicists since Lord Kelvin used it to estimate the age of the earth (Thomson, 1862), it has not been exploited as much as it might in the accelerating search for contemporary climatic change.

It is now well-known that human activities are changing the composition of the atmosphere and this, in turn, is likely to cause substantial changes in global climate. Because of the great social, economic, and scientific importance of such changes, an intensive search is underway for reliable physical models to predict its course and for reliable data bases to confirm the predictions (Ellsaesser *et al.*, 1986). Existing models indicate that by the middle of the next century, mean global temperatures will increase a few degrees Celsius with effects in the Arctic being greatly accentuated (Dickinson, 1986). The most studied database is the surface air temperature, which shows large variability in time and space on all scales. Few records in the Arctic extend back far enough to put the effects of the modern buildup of greenhouse gases in the context of climatic changes known to occur on longer time scales.

Unlike air temperatures, which depend on a thermal environment that changes hourly, temperatures beneath the surface in cold permafrost represent a systematic running mean of the recent temperature history at the surface; they are unaffected by thermal complications from moving fluids. Because a temperature change at the surface takes time to propagate downward into the earth,

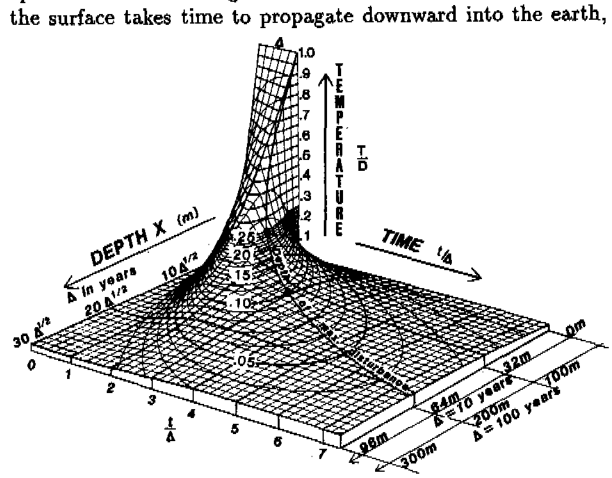

Figure 1: The thermal memory of the earth for events on its surface. Contours show the strength of the temperature anomaly at any time t and depth x due to a uniform surface temperature disturbance of magnitude D and duration Δ. Depth axes give depth in meters for events with durations of 1 century or 1 decade (on the right) or for an arbitrary "Δ" years (on the left). (Assumed thermal diffusivity is $10^{-6}m^2sec^{-1}$.) Two event-durations after the conclusion of an event (*i.e.*, at $t = 3\Delta$), the maximum signal falls to 10%, and it occurs at about 38 m for $\Delta = 10$ years and 120 m for $\Delta = 1$ century. At the conclusion of a uniform disturbance of duration Δ (*i.e.*, at $t/\Delta = 1$), the anomaly is not appreciable (<5%) below about 50 m for $\Delta = 1$ decade, and about 150 m for $\Delta = 1$ century.

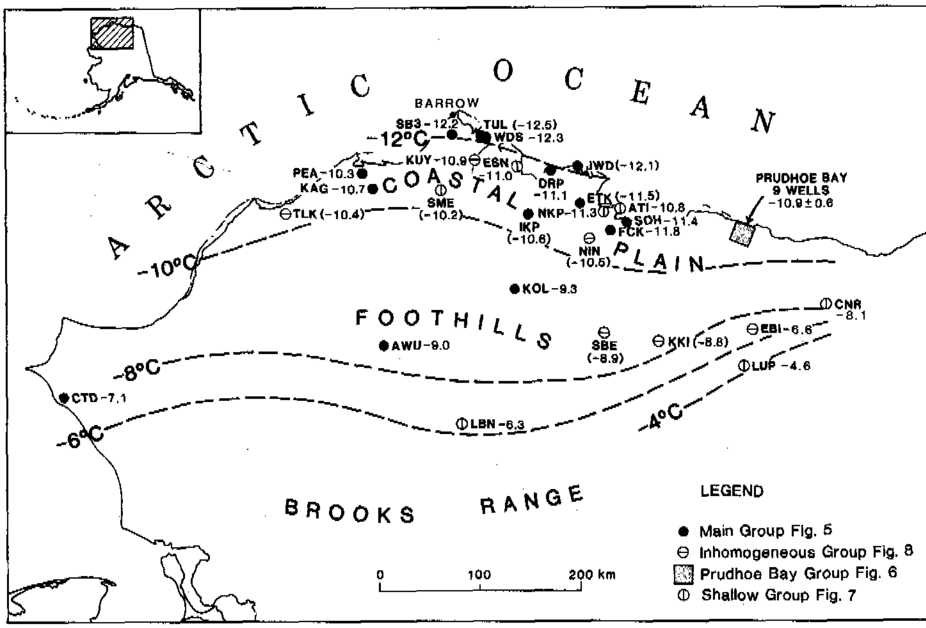

Figure 2: Map of northern Alaska showing well locations with symbols keyed for Figs. 5 to 8. All except the "shallow group" indicate surface warming over a period of several decades to a century. Numbers and contours (dashed curves) represent the extrapolated long-term mean annual temperature at the top of permafrost, θ_o.

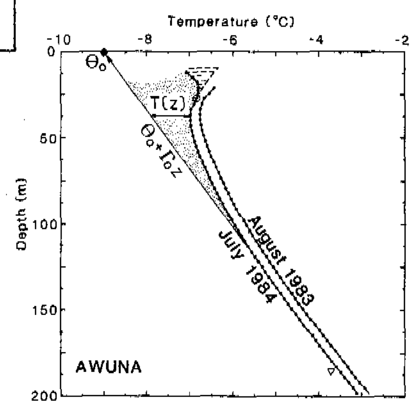

Figure 3: Measured profiles at AWU. Line with slope Γ_o and surface intercept θ_o is least-squares fit to the linear portion. $T(z)$ (stippled region) is temperature anomaly identified with secular increase in surface temperature; cross-hatched region indicates recent cooling. Dots show every tenth measurement point; their diameter ($\sim$0.05°C) is estimated reproducibility.

the deeper the temperature measurement, the farther back in time is the interval of surface temperature history it represents, and the smoother is the signal as the high-frequency noise is progressively filtered out. Thus the ground "remembers" the major surface temperature events (Birch, 1948), and careful temperature measurements made today to depths of a few hundred meters can provide information on the history of local surface temperature during the past few centuries (Figure 1).

We have recently applied these methods to show that permafrost has been warming rapidly in the 20th century over much of a 10^5 km^2 region of the Alaskan Arctic (Lachenbruch and Marshall, 1986; for related discussion, see Judge and Pilon, 1984; MacKay, 1975; Brown and Andrews, 1982). We shall reiterate and update these results and discuss them in the context of new climatological information on century-scale trends in Arctic air temperature.

OVERVIEW OF THE GEOTHERMAL REGIME

The data for this study represent typical conditions in cold continuous permafrost (*i.e.*, permafrost with temperatures well below 0°C and no mobile liquid fraction). They were obtained from oil exploration wells distributed throughout an area of rolling foothills and lake-strewn coastal plain between Alaska's Brooks Range and the Arctic Ocean (Figure 2). The major features of the temperature observations and their interpretation in terms of heat-conduction models are illustrated by the data from Awuna, a site in the Foothills province 150 km from the Arctic Coast (AWU, Figure 2). The two latest profiles, taken 2.4 and 3.3 years after drilling, are shown in Figure 3; their difference represents the dissipation of the heat introduced by drilling. A conduction model of the drilling process (Lachenbruch and Brewer, 1959) shows that at this late stage, the disturbance ($\sim$ 1°C) is the same at all depths and the *form* of the profile is no longer disturbed. The linear portion of the profile at depth in Figure 3 represents steady flow of heat q from the earth's interior according to the Fourier relation

$$q = K\Gamma_0 \qquad (1)$$

where Γ_0 is the thermal gradient and K is thermal conductivity.

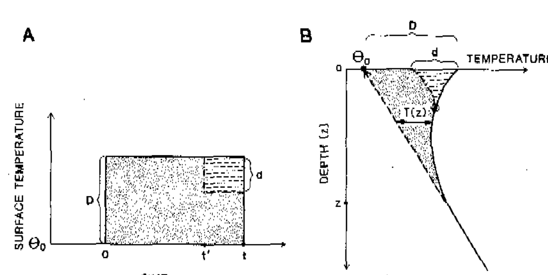

Figure 4: Schematic representation of the effect of surface temperature history on the geotherm. A step increase in surface temperature in the amount D between time 0 and t [solid rectangle in (A)] results in departure of the upper part of the geotherm [solid curve in (B)] by $T(z)$ from the steady-state linear geotherm [dashed line in (B)]. A superimposed cooling in the amount d between t' and t [dash-dot line in (A)] would modify the geotherm as shown by the dash-dot curve above the circled point in (B).

Extrapolation of the linear portion to the surface yields the intercept temperature θ_0. Our main interest is in the upper 100 m or so and the curvature there that causes a warming departure of the geotherm from the extrapolated linear profile; the resulting temperature anomaly (stippled area in Figure 3) is denoted by $T(z)$. In the simplest interpretation, $T(z)$ represents the response of the ground to recent warming of the mean annual surface temperature from a previous long-term value of θ_o; the gradient reversal above the circled dot could represent a very recent cooling. This interpretation is illustrated in Figure 4B for a simple hypothetical temperature history shown in Figure 4A. [The theory of such changes has been widely discussed (*e.g.*, Birch, 1948; Cermak, 1971; Osterkamp, 1984).]

The configuration of the geotherm at Awuna is not unique (Figure 5). In fact, Figures 5 to 8 (representing all stations shown in Figure 2) show that most sites display a similar warming anomaly $T(z)$ (stippled areas), although many show a more pronounced cooling anomaly than AWU superimposed in the upper layers (cross hatched). Irregularities can make the anomaly difficult to evaluate or, in extreme cases (Figure 8), to identify. In Figures 5 to 8, the curves are offset to display their form; consequently, the intercept value at the surface, θ_o, does not appear. The individually estimated values of θ_o are shown and contoured in Figure 2. The fact that they can be contoured in this simple manner—values increase systematically with distance from the coast in a pattern modified by regional topography—supports the interpretation of θ_o as a long-term mean surface temperature with regional climatic significance.

THE SIMPLE MODEL

The simple representation of Figure 4 implies that curvature in the geotherm is caused by changing temperature at the earth's surface. Although cold permafrost permits us to neglect heat transport by moving ground water and treat this as a problem in pure heat conduction, three other sources of curvature are theoretically possible: 1) transient effects of rapid aggradation of the surface (if the boundary is not static); 2) steady-state effects of lateral heat flow (if the problem is not one-dimensional), and 3) steady-state effects of systematic changes in thermal conductivity with depth (if the medium is not homogeneous).

In the first mechanism, the gradient diminishes near the surface because the rapidly accumulating sediments don't have time to warm up. To explain the observed curvature, however, would require persistent sedimentation rates on the order of 10 cm yr^{-1} or more, a condition that does not occur at the Alaskan sites. The three-dimensional effects of the second mechanism do exist in this terrain (*e.g.*, Lachenbruch *et al.*, 1962); surface temperature varies laterally primarily according to geomorphic control of the moisture regime. These effects must contribute to the anomaly $T(z)$ and add to the scatter of estimates for climatic change based on $T(z)$. Nevertheless, the uniformity of sign of the observed $T(z)$, generally a warming, and the contourability of θ_o (Fig. 2) suggest that three-dimensional effects are secondary; the wells were not drilled selectively in warm places such as lakes or topographic highs (see Fig. 10, Lachenbruch and Marshall, 1986).

The third mechanism, vertically varying conductivity, causes the greatest problem for the simple model although it is clear that the warm regions (stippled areas) are not caused exclusively by conductivity contrasts. No evidence exists for the systematic increase in conductivity required to explain the systematic decrease in gradient in the upper 100 m. In wells like Awuna (Figure 3) and most of those illustrated in Figures 5 and 6, the uniformity

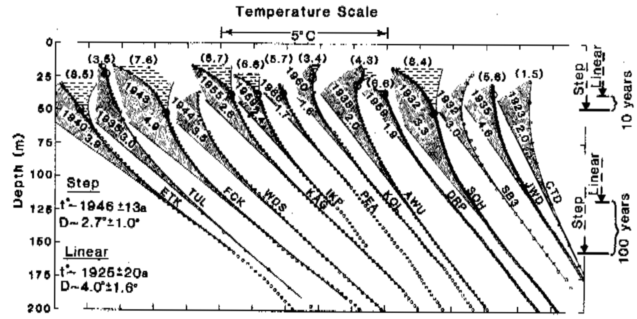

Figure 5: Data from main group (Fig. 2). Stippled region is our interpretation of warming anomaly $T(z)$. Temperature change (°C) and date (A.D.) are given for the best-fitting step-function climate history (solid curve). Cross-hatched area above circled dot represents recent cooling. Number in parenthesis is age of construction pad in years at time of temperature observation. Summary statistics are for best-fitting step and linear temperature histories. Thermal diffusivity assumed to be $10^{-6}m^2sec^{-1}$ except for measured value ($1.4 \times 10^{-6}m^2sec^{-1}$) at CTD.

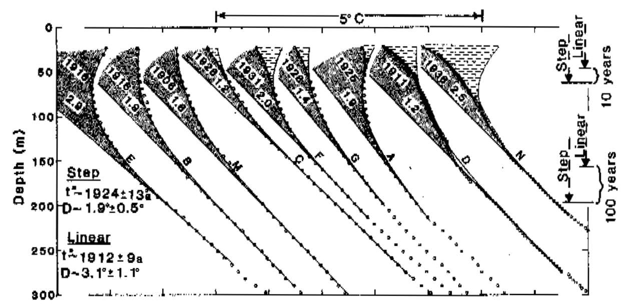

Figure 6: Data from Prudhoe Bay group (Fig. 2). Assumed thermal diffusivity is $1.6 \times 10^{-6}m^2sec^{-1}$ based on measurements by Lachenbruch *et al.* (1982). Pad ages not determined. Other conventions as in Fig. 5.

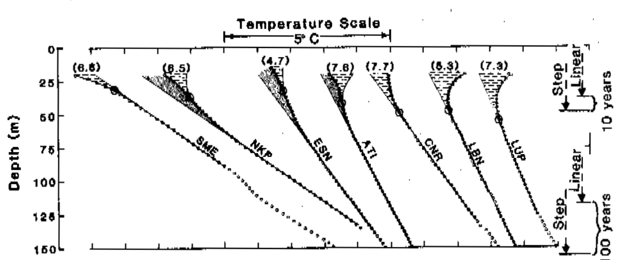

Figure 7: Data from shallow group (Fig. 2). Cross-hatched region represents most recent event. Other conventions as in Fig. 5.

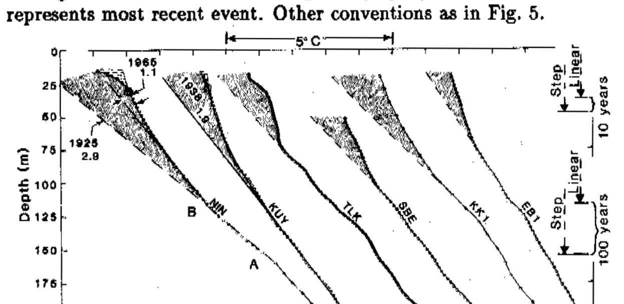

Figure 8: Data from inhomogeneous group (Fig. 2). Gradient changes at depth (*e.g.*, at "A") are caused by conductivity contrasts. Gradient change at B in NIN is ambiguous; if caused by conductivity contrast, the best-fitting step-history is a warming of about 1°C in the 1960's; if not, it is about 3°C in the 1920's. Conventions as in Fig. 5.

of gradient at depth permits linear extrapolation to the surface with reasonable confidence that K is uniform and therefore $T(z)$ is truly a transient anomaly. Profiles with less uniform gradients indicate contrasts in conductivity, and for them it is often difficult, in the selection of Γ_o, to distinguish between slope changes caused by climatic change and those caused by conductivity change (see for example, NIN in Figure 8). Gradual slope change at greater depth caused by conductivity changes could be confused with older climatic events. These ambiguities could be resolved if the thermal properties were known as a function of depth, but we do not have such information for most of the wells. Nevertheless, it seems clear from inspection of the profiles (Figures 5 to 8) and from the foregoing discussion that the consistent departure from linearity toward warmer temperatures generally represents a transient warming. If the gradient actually reverses sign (for example, at 45 m in Awuna), transient warming is the only plausible explanation.

Even if the foregoing geometric conditions for the simple model are satisfied, the warming anomalies need not all be associated with changing climate; locally they can be caused by engineering modifications or natural geomorphic processes on the tundra surface. In general, the engineering disturbances will not pre-date the construction for drilling at these wilderness sites, but natural growth and deterioration of thaw lakes and bogs and the shifting of stream channels and shorelines can be expected to introduce local variability into the general temperature patterns established by climate. Such local anomalies are actually of great interest in themselves; they contain information on the long-term passive effects of engineering structures and on the evolution of geomorphic features (Gold and Lachenbruch, 1973).

We note finally that the analysis of curvature in the simple model depends upon identifying an undisturbed straight-line portion of the geotherm at depth (Figure 3). Given the variability of climate, there is, of course, no such thing as an "undisturbed geotherm" even "at depth." However, plausible perturbations that occur much farther in the past than the multi-decade events we are analyzing generally have relatively small effects on geothermal gradient, and their effects on curvature are practically undetectable. For example, the "little ice age" with a temperature anomaly of -0.5°C between 1550 and 1850 A.D. can be seen from Figure 1 (Δ = 300 yrs, $t \approx 1.5\ \Delta$, D= -0.5°C) to cause a maximum present-day anomaly of -0.125°C at about 125 m, an average gradient perturbation of 1°C/km. Gradient effects of late Wisconsin glaciation (*e.g.*, Δ = 30,000 yrs, $t \approx 1.5\Delta$, D = -5.0°C, Figure 1) are of comparable magnitude. Such effects have little influence on our analysis of the large recent perturbations such as that at Awuna (Figure 3) which dominates (*i.e.*, reverses) the local geothermal gradient of 30°C/km.

It is clear that inhomogeneities and three-dimensional effects limit our capacity to resolve details of $T(z)$, and the lack of samples requires us to estimate thermal conductivity and make some arbitrary choices for the "undisturbed profile at depth." Nevertheless, simple direct calculations based on the homogeneous one-dimensional transient model discussed provide insight into time scales and the general magnitude of recent temperature changes involved in generating the observed anomalies. For this purpose, we write the temperature (Figure 3)

$$\theta(z,t) = \theta_o + \Gamma_o z + T(z,t) \tag{2}$$

The last term in (2) is governed by the differential equation

$$\frac{\partial^2 T}{\partial z^2} = \frac{1}{\alpha}\frac{\partial T}{\partial t} \tag{3}$$

where $\alpha = K/\rho c$ is thermal diffusivity, and ρc is volumetric heat capacity.

We assume that the surface temperature varies with time according to a simple three-parameter power law

$$T(0,t) = D\left(\frac{t}{t^*}\right)^{n/2} \qquad \text{for } 0 < t \leq t^* \quad n = 0,1,2,\ldots \tag{4}$$

This provides the surface boundary condition for the solution of equation (3). The initial condition is

$$T(z,t) = 0 \tag{5}$$

The solution of equations (3), (4), and (5) evaluated at $t = t^*$ is (Lachenbruch *et al.*, 1982; Carslaw and Jaeger, 1959)

$$T(z) = D2^n\Gamma\left(\frac{1}{2}n+1\right) i^n\mathrm{erfc}\frac{z}{\sqrt{4\alpha t^*}} \tag{6}$$

where $i^n\mathrm{erfc}\beta$ is the n^{th} integral of the error function of β and $\Gamma(\beta)$ is the gamma function of argument β. This result gives the ground temperature at the end of a surface warming interval of duration t^* during which the surface temperature increased by D. The form of the increase can be adjusted with the value n; for example $n = 0$ is a step increase in temperature, and $n = 2$ is a linear increase (equation 4).

APPLICATION OF THE SIMPLE MODEL AND IMPLICATIONS FOR LONG-TERM HEAT BALANCE

An example of the best fitting results for Awuna for several values of n in the simple model (equation 6) is shown in Figure 9B. If the warming occurred as a step function ($n = 0$), it started about 50 years ago and warmed about 2°C; if it was a linear change ($n = 2$), the increase was 2.6°C during the last 68 years; an increase accelerating as the square of time ($n = 4$) would have started 97 years ago and grown to 2.9°C. Figure 9A shows that all four of the surface temperature histories illustrated in Figure 9B fit the observations equally well, generally within the overall measurement error in the interval below 30 m where they were fit. The shallower observations (above the circled dot in Figure 3) were not included because they represent a recent complication incompatible with the simple form of equation (4); they will be considered in the next section. Although the Awuna example illustrates the ambiguity of historical details in this inverse method, it leaves little doubt that we are looking at the effects of a multidecade event in the last century that represents a temperature increase of 2°or 3°C.

This warming results from a net accumulation of heat in the earth at a rate which we can now calculate. Average values are 0.1–0.2 watts per square meter (climate flux C, Figure 10); the steady geothermal flux flowing in the opposite direction (g, Figure 10) is about 0.06 Wm^{-2}. It is of interest to compare these gentle solid-earth fluxes to the more vigorous activity on the other side of the earth's surface. The larger "top-side" fluxes drive the ecosystem, and their balance point determines the earth's surface temperature (Weller and Holmgren, 1974). They consist of the incoming and outgoing radiation components and their difference ("net," Figure 10A) which, among other things, is responsible for melting snow and evaporating water allowing them to return to the sea or sky to balance both the thermal and hydrologic budgets in preparation for the next annual cycle. The numbers on the arrows (Figure 10A) indicate that these top-side climatic fluxes sum to zero as they should if the climate is not changing. But we just determined from the solid earth (Figure 9) that climate *is* changing; at Awuna about 0.16 Wm^{-2} more has been going into the earth than out. As this is just 1/100th of the net radiation (itself a difference of large numbers), we could never detect it by trying to keep a balance sheet of fluxes at the earth's surface. Thus while the unbalanced climatic flux (C, Figure 10A) is an inconspicuous second-order effect in the climatic system, it is a conspicuous first-order effect that dominates the thermal regime of the upper 100 m in the solid-earth system (Figure 10B)—the solid earth is a good monitor for changes in the surface energy balance. The downward flowing heat cannot go far in a century because the earth is a poor conductor; it is all contained in the stippled region (Figure 10B), which represents the complete climatic event from start to finish.

The analysis illustrated for Awuna (Figure 9) has been extended to the other sites where curves were smooth enough to permit its application (Figures 5, 6, 7). In performing the analysis, we used an iterative least-squares technique to determined the best-fitting parameters D and t^*, equation 6, for the step ($n = 0$) and linear ($n = 2$) climatic change after identifying the long-term anomaly $T(z)$. An error in the choice of the "undisturbed" linear profile at depth ($\theta_o + \Gamma_o z$) can make a big difference in the outcome (see NIN, Figure 8). Where it was ambiguous we generally opted for

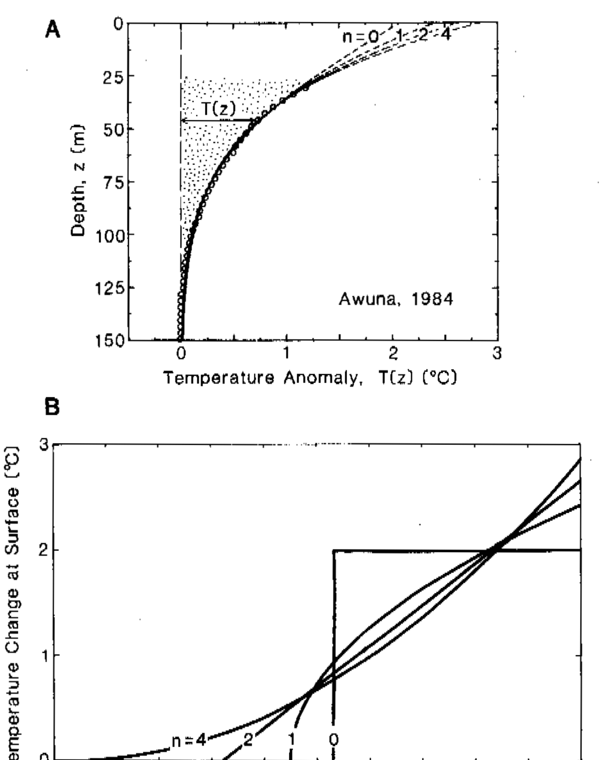

Figure 9: (A) Warming anomaly $T(z)$ obtained from AWU (Fig. 3). Circles are observations; curves are best-fitting geotherms calculated by Eq. 6 for the four forms of temperature history (Eq. 4) shown in (B).

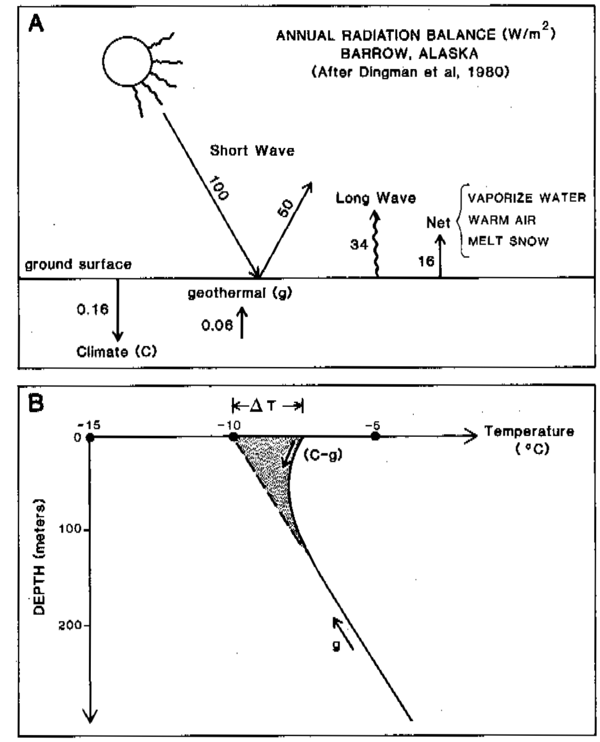

Figure 10: (A) Comparison of typical unbalanced climate flux, C, with geothermal flux, g, and other fluxes at earth's surface. (B) Role of climate flux in the geothermal regime.

the choice that would give the lesser age for the event. In Figures 5 and 6 the solid curves joining the observations, and the numbers within the stippled regions, are for the step model; also shown are statistical summaries for both the step ($n = 0$) and linear ($n = 2$) models. In the right-hand margins of these figures, arrows show the depths at which effects of step and linear changes in surface temperature, in progress for 10 or 100 years, generally become small (<5% of their surface value; see Figure 1). It is clear from this reference alone that the duration of warming was much more than a decade but not much more than a century. The formal summaries for the Prudhoe Bay group (Figure 6) indicate an "average" linear change of about 3°C starting in 1912, or an "average" step change of about 2°C starting in 1924. For the main group (Figure 5) the corresponding values are a linear change of about 4°C starting in 1925 or a step change of 2.7°C starting in 1946.

Unlike the closely spaced Prudhoe Bay group, statistics for the main group do not represent a geographic area because we have excluded those sites (shown in Figure 7) having no evidence of a temperature anomaly extending deeper than 50 or 60 m. Three of the sites, NKP, ESN, and ATI, show a warming (below the circled points on profiles), which suggests a start-up between one and two decades ago when we use the previous assumptions. Like many others, these wells show a recent transient cooling trend (above the circled dots). The profile at SME is unique for the Coastal Plain and Foothills provinces in showing only the recent cooling above the circled point; a previous warming trend is not indicated. The remaining three sites, CNR, LBN, and LUP, form a distinct group lying along the front of the Brooks Ranges (Figure 2). They show only a superficial warming trend (above the circled point) that is largely a product of the last decade.

Consideration of Figures 5 to 8 collectively, makes it clear that although our estimates for temperature change in the last century contain large uncertainties and the derived parameters do not show spatial trends that can be readily contoured, it would not be correct to treat these as random variations in the description of a uniform climatic event. The high quality of the data in Figure 7 precludes the possibility of long-term warming indicated at most of the other sites, and the warming at Prudhoe Bay probably started earlier than in some areas farther west and inland. The most reasonable conclusion is that although there was a fairly consistent warming trend of a few degrees Celsius at the permafrost surface during the last century in the Coastal Plain and Foothills provinces, it did not start at the same time throughout this region and probably did not extend into the Brooks Range.

EVENTS OF THE LAST DECADE

In the foregoing analyses, the curves that we fit to the lower older part of the geothermal anomaly often do not fit very well in the upper 50 m or so, *i.e.*, above the circled dots. Judging from the marginal arrows in Figures 5–7, the departure (cross hatched region) results largely from thermal events of the decade preceding the last measurement. This is confirmed by a succession of measurements at KAG (Figure 11) where we were able to observe the evolution of such an anomaly. The two curves in Figure 11A illustrate three transient processes in progress simultaneously: 1) the dissipation of the drilling disturbance from 1979 to 1984 without changing the form of the geotherm, 2) a 45-year long-term

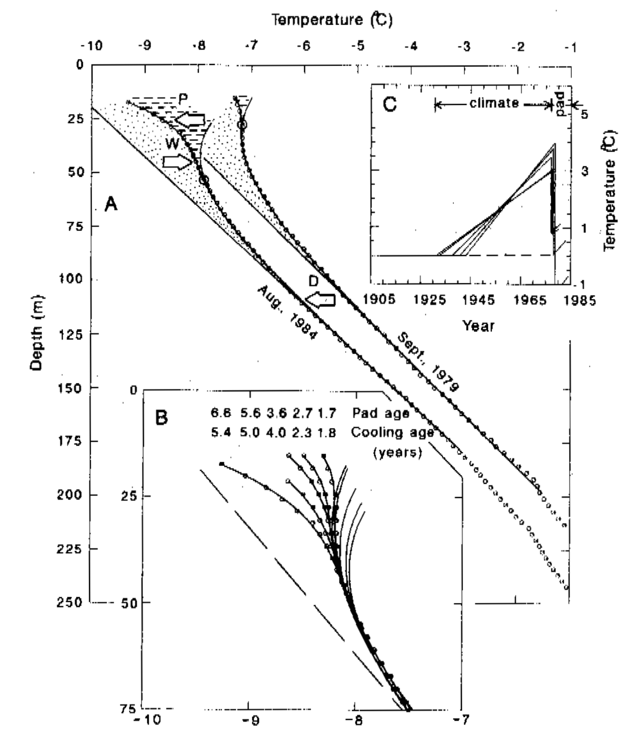

Figure 11: (A) Two profiles from KAG showing thermal transients caused by: dissipation of drilling disturbance (D), climatic warming (W), and cooling since pad construction, P. (B) Superimposed measurements showing evolution of recent cooling anomaly. Solid curves correspond to surface histories shown in C. Numbers in B compare age of pad at time of measurement with duration of cooling as determined from independently calculated surface histories shown in C.

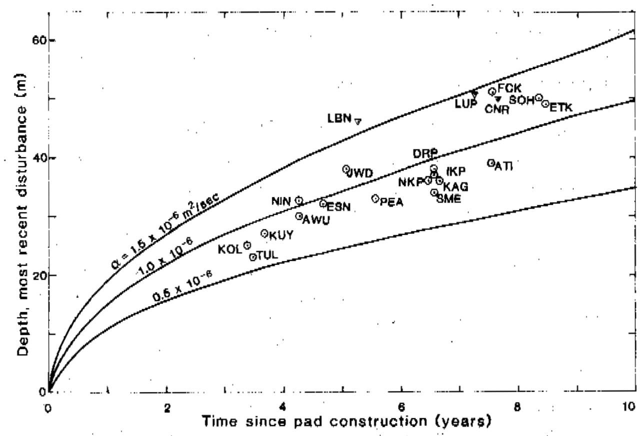

Figure 12: Estimated depth of penetration of most recent thermal disturbance (depth of circled dots, Figs. 5, 7, and 8) *versus* age of construction pad at time of last observation. Curves show depths to which a step change in surface temperature at time of construction would penetrate for a range of thermal diffusivities. Circles indicate a cooling disturbance; triangles, a warming disturbance.

warming (stippled) approximated by the linear trend ($n = 2$), and 3) a cooling event (cross-hatched) propagating downward and causing a growing departure between the geotherm and the theoretically fit curve. The development of the last anomaly can be illustrated by superimposing the two sets of measurements and combining them with three intermediate sets (Figure 11B). For each of these five curves, an independent climatic history was determined by fitting a linear trend ($n = 2$) to the deep part, and a superimposed step ($n = 0$) to account for the departure near the surface. The resulting histories (Figure 11C) all indicate that the cooling started at about the same time, which corresponds to the date of drilling-pad construction. This redundancy provides a basis for confidence in the conduction models and the choice of thermal diffusivity ($10^{-6}\mathrm{m}^2\mathrm{sec}^{-1}$). It also suggests that at least part of the widespread recent cooling might be the passive thermal disturbance caused by emplacing a gravel construction pad on the natural tundra surface. Indeed, it is seen from Figures 5 and 7 that even though almost all profiles were taken in the same month (8/84), the depth of the most recent disturbance (circled dot) generally increases with the age of the pad (numbers in parentheses). Figure 12 shows that the relation between these two quantities is consistent with a step temperature change at the time of pad construction for a reasonable range of thermal diffusivities.

In the Coastal Plain and Foothills provinces, this recent trend is generally a cooling on the order of a degree or more; in some cases it nullifies much of the climatic warming of the last century. Although we do not know the cause, we speculate that the slightly elevated, smooth, unvegetated construction pad might have a thinner drifted snow pack and less insulation from the winter cold than the surrounding terrain. Such effects can lower the mean ground temperature significantly (Goodrich, 1982). However, at the three sites (LBN, LUP, and CNR, Figures 2 and 7) at the edge of the Brooks Range province, the recent event is a warming and at Prudhoe Bay we do not yet know whether there are any systematic engineering effects. Engineering practices have varied widely at these sites and much more study is needed to document their history and to determine their long-term effects on the heat balance of the tundra surface. For the purpose of the present discussion, however, the observations underscore the sensitivity of the permafrost temperature to modifications of the surface and the need to control these modifications near any observation wells that might be drilled to monitor long-term natural temperature changes.

RELATION TO WEATHER RECORDS

We have described a method of estimating the temperature history of the permafrost surface in the recent past from a knowledge of the geotherm today. Except for a recent engineering disturbance and an overlay of random geomorphic effects, we believe that the results reflect a history of changing climate. The most familiar indicator of changing climate is the history of surface air temperature and it is instructive to compare our results to the best estimates of this quantity for the same region.

The only weather record of consequence on the Alaskan North Slope is at Barrow where observations began in 1921 (Figure 13A). However, in a comprehensive analysis of world-wide records, Hansen and Lebedeff (1987) have recently shown that air temperature changes at mid- and high-latitude stations are highly correlated if the station separation is less than 1000 km. On this basis, they combined station data to describe long-term changes for all regions of the earth's surface. Results from their paper are shown as parts C and D of Figure 13, and from their computer tape (which they kindly made available) as part B.

Figure 13 shows clearly that the warming of the last century is greater at high latitudes, and additionally, that it is greater in the North American quadrant of the Arctic than in the Arctic as a whole. The only record in our area of study, Barrow (Figure 13A) shows a decrease over the period of observation as would other generalized Arctic records over the same period.

To close the gap between the climatological and geothermal estimates of surface temperatures, we suppose curve B to be a reasonable representation of air temperature, and ask what the corresponding geotherm would look like if curve B also represented temperature at the top of permafrost. The results of the calculation (Figure 14) show clearly that 100 years of weather records are not enough to answer the question. An additional assumption must be made for the pre-1880 long-term mean temperature (θ_o in our models). For Case I, it is taken to be the start of the regression line for 1880 to 1925, θ_o', in Figure 13B; for case II it is the start of the regression line for 1880 to 1985, θ_o'' in Figure 13B, and for case III we assume it to be the same as that for the last 65 years, the period of the Barrow records (θ_o''', Figure 14B). The ambiguity illustrated by the three cases shows that the geotherm contains a great deal more information on past trends in climate than 100 years of weather records; the extra information is provided by the geothermally identified long-term mean (θ_o).

A second interesting feature of Figure 14 is the comparison of magnitudes and depths of the climatic anomaly for the "synthetic" geotherms I, II, and III, and the geotherm for Awuna (and similar ones in Figures 5 and 6). It shows: 1) the air temperature changes are large enough to account for the geothermal observations under reasonable assumptions for the long-term mean (*e.g.*, cases I or II, Figure 14); 2) the contradiction between case III and measured geotherms (Figures 5 and 6) proves the warming trend of the last 100 years is substantially more than the return from a deep minimum in the late 19th century, and 3) the synthetic geothermal anomalies extend deeper than that for AWU and others in Figure 5, (they would also be deeper than those in Fig. 6 for similar assumed diffusivity) implying that the atmospheric warming started earlier than the permafrost warming.

To continue the numerical experiment, we test our method of estimating surface temperature history by fitting effects of simple

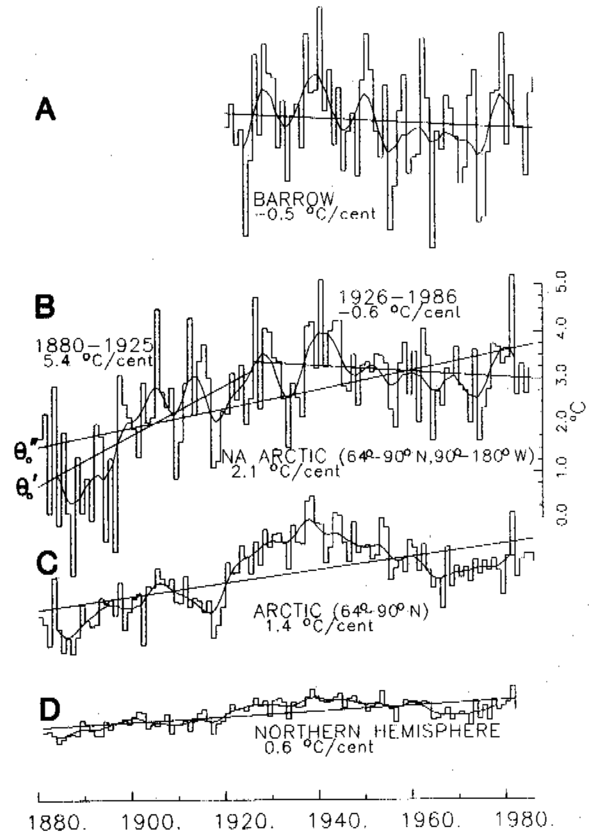

Figure 13: Mean annual temperatures from (A) U.S. Weather Bureau at Barrow, Alaska, and (B) from Hansen and Lebedeff (1987) for the North American quadrant of the Arctic; (C) the circumpolar Arctic, and (D) the Northern Hemisphere. Curves for nine-year weighted average and selected regression lines also shown.

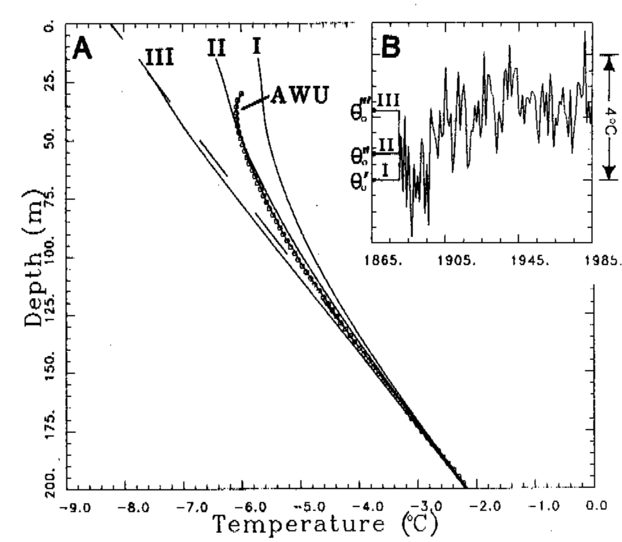

Figure 14: (A) Synthetic geotherms based on following assumptions: geothermal gradient and diffusivity same as AWU, surface temperatures given by inset (from 13B), and long-term mean temperature of θ_o' (case I), θ_o'' (case II) or θ_o''' (case III).

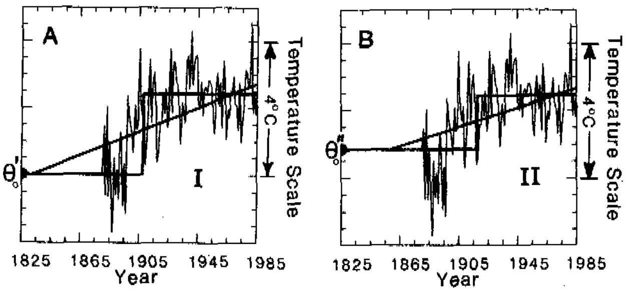

Figure 15: Step and linear reconstruction of climatic history by our standard curve-fitting technique applied to synthetic geotherms I (in A) and II (in B) from Fig. 14.

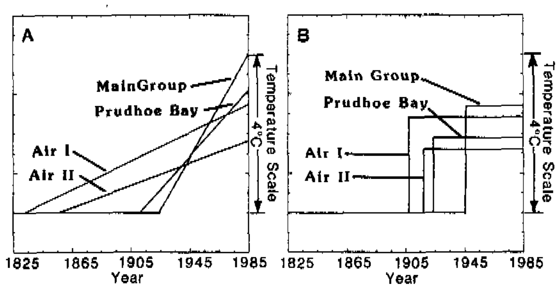

Figure 16: Comparison of average surface temperature histories for the linear (A) and step (B) fits to geotherms from the main group (Fig. 5), Prudhoe Bay group (Fig. 6), and synthetic geotherms of Fig. 15A (Air I) and 15B (Air II).

(*i.e.*, step and linear) changes to the synthetic geotherms (I and II, Figure 14) as we did previously for AWU in Figure 9. Figure 15 shows that these standard techniques do indeed give reasonable approximations to the original driving function which for the geothermal case, of course, is unknown.

To summarize these comparisons, we show in Figure 16A the linear surface temperature trends (Figure 15) that we deduced from the fictitious geotherms (I and II, Figure 14) along with the corresponding "average" values deduced from the geotherms in the main group (Figure 5) and Prudhoe Bay group (Figure 6). Results for the corresponding step approximations are shown in Figure 16B. Figure 16 provides a common basis for comparison of highly generalized average trends for surface temperature; one derived from measurements in the air in a large ($\sim 10^7$ km^2) Arctic region, the other from measurements in the solid earth in a sub-region ($\sim 10^5$ km^2). We have been surprised to learn that these generalized air temperature changes are essentially as large as our geothermal estimates for the surface of permafrost in the Alaskan Arctic although the permafrost changes seem to be somewhat more recent. Finally, it is important to bear in mind that the geotherms show that at some of the permafrost stations significant trends have *not* occurred in the last century; perhaps the same is true at some of the climatic stations.

DISCUSSION

Anomalous curvature in temperature profiles from oil wells indicates a 20th century warming trend of 2°–4° C at the surface of permafrost over much of a 10^5 km^2 region of the Alaskan Arctic. The result comes from analysis of precision temperature measurements by heat-conduction theory which can be applied with confidence because ground water cannot move in the cold continuous permafrost. Although the long-term warming to 100 m shown by these curves is widespread in the Alaskan Arctic, it is not universal there. Counter examples occur at the points in the "shallow group" of Figure 2 (7 out of 36 sites). Other complications exist in the upper 50 m from engineering disturbance, geomorphic processes, and short-term climatic events of the last decade. We do not have enough information to map the patterns of the long-term trends or to understand their causes. It is clear, however, that the presence of these systematic transient temperature anomalies in the upper 100 m of permafrost is a firm indication that the heat balance at the surface was systematically disturbed during the last century, and where such signals are absent, it is a clear indication that it was not. This geothermal signal is a direct thermophysical consequence of the surface thermal events we are looking for, not a "proxy" for them. It is possible to obtain climatic history information for the last century "after the fact" by drilling a 200 m observation well at any location where the information is desired. For example, where a changing tundra ecosystem was under study, such a well could provide site-specific information on temperature history of the surface over the past two centuries or so. An additional advantage of the geothermal method is that the earth integrates the permafrost temperature uniformly and continuously, filtering out the (high frequency) noise and preserving only the (low frequency) climatic signal. This spares us a computational burden, and avoids uncertainties that might be caused by changing measurement conditions during the accumulation of weather records over many human generations.

A peculiarity of the geothermal method is that the "surface" whose temperature is being reconstructed is not strictly the surface of the solid earth. It is the surface of permafrost (T_{pf}, Figure 17) which lies beneath an annually thawing "active layer" whose thickness varies from 0.2 m in wet areas to 2 m in dry ones. Below this surface, heat transfer is almost exclusively by heat conduction, above it is not. Most climatological studies of contemporary change are analyses of T_{air} (Figure 17) measured in a standard observatory thermometer shelter. Between it and the top of permafrost are a boundary layer in the air, a seasonal snow pack and the active

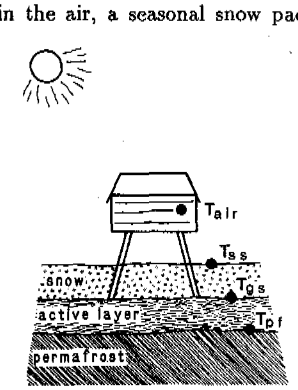

Figure 17: Measurement sites for the different annual mean temperatures: T_{pf} at upper surface of permafrost, T_{gs} at the ground surface, T_{ss} at the solid surface of the snow pack when it is present and ground surface when it is not, and T_{air} in a standard observatory thermometer shelter.

layer, each of which transfers heat and moisture in complex ways, and each can be influenced by changing geomorphic conditions. Thus it might not be surprising that weather records at Barrow suggest that T_{air} did not change systematically in the last 65 years (Figure 13A) but nearby geothermal measurements suggest that T_{pf} probably did (SB3, Figures 2 and 5). The difference could relate to a variety of factors (Smith and Riseborough, 1983) including, of course, secular change in snow cover or other moisture conditions, important climatic variables in any case. To understand and monitor the presently changing arctic climate, it is important to understand the physical relations between temperatures at the top of permafrost and those routinely measured in the overlying air so that geothermally detected changes can be more directly related to the wealth of data on air temperature change. It is interesting that statistical treatment of 100 years of sparse Arctic weather records (Hansen and Lebedeff, 1987) suggest a total warming of air temperature of the same general magnitude (although not quite as recent) as our estimates for warming of the permafrost surface.

A second obvious and important area for study is to determine how widespread the geothermal climatic effects are and what they tell us about the pattern of contemporary change (both in permafrost and in other impermeable formations worldwide). References to such change, mostly in site-specific local studies scattered throughout the geothermal literature, need to be pulled together and systematic additional measurements should be undertaken to understand existing conditions and to monitor their future changes.

Finally, our data indicate that the permafrost surface in the Alaskan Arctic is generally warming rapidly and this implies a rapid change in climate there. Why is this happening? Perhaps it results from a change in snow albedo due to the settling out of atmospheric pollutants (*e.g.*, Clarke and Noone, 1985), or possibly to a secular change in sea ice cover over the nearby Arctic Ocean or to subtle changes in other parts of the climate system of the North Polar Basin (Weller, 1984; Barry, 1985). Whatever its cause may be, however, it is important that it be understood and considered in the context of predicted global change. The permafrost warming is a very measurable and conspicuous effect occurring where the global warming is expected to be greatest and first observable and in a medium (continuous permafrost) well-suited to detection and monitoring of such changes. It is important to understand the changing permafrost regime whether it is a direct effect of predicted global change or a complicating background effect upon which the predicted global change will eventually be superimposed.

Acknowledgments

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THE SIGNIFICANCE OF CLIMATIC CHANGE FOR THE PERMAFROST ENVIRONMENT

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ABSTRACT The nature and magnitude of climatic changes that are predicted in association with the greenhouse effect are considered in terms of their implications for the permafrost environment. With a climatic warming it is inevitable that some permafrost would eventually disappear and ultimately there would be a new "permafrost map". This would be a slow process, however, but a more important and immediate question is how quickly degradation effects would be seen, and at what rate would they develop. The relationship between climate (change) and permafrost is not a simple one, and the possible modulations introduced by local (microclimatic) and lithologic conditions are illustrated. Finally, the general implications of climatic change for periglacial processes are also briefly considered.

INTRODUCTION

There is growing interest in the impact of human activity on the levels of CO_2 and other radiatively active trace gases in the atmosphere, and their likely effect on climate. Climatologists have shown that air temperatures would increase because of the so-called "greenhouse effect", although it is not clear by how much. Studies demonstrate, however, almost universal agreement on a warming concentrated in the polar regions, along with increased precipitation there; for a doubling of the atmospheric CO_2 content, some models predict that annual air temperatures throughout the Arctic could increase by 3 to 6C, and perhaps even more in some areas (Harvey 1982, Hansen et al. 1984).

While a climatic warming trend has yet to be clearly identified in the meteorological record (e.g. see Weller 1984), the magnitude of such changes in climate would produce serious and far-reaching environmental and engineering problems in permafrost regions, and for the arctic environment as a whole (McBeath 1984, French 1986). Historical evidence shows that while the northern hemispherical temperature increased by only about 0.5C from 1880 to 1950, temperatures in the upper two metres of permafrost generally increased about 2C or more during the same period throughout much of northernmost Alaska, according to calculations of Lachenbruch and Marshall (1986).

CLIMATIC CHANGE AND THE PERMAFROST ENVIRONMENT

Permafrost is unique in earth material terms, since it exists close to its melting point. Most discontinuous permafrost is either relict, or in such delicate balance that climatic or other environmental changes can have drastic disequilibrium effects. Tens of thousands of square kilometers of permafrost are warmer than -3C, and we can expect that most of it would eventually disappear under the climatic warming predicted, although complete degradation would take many centuries, at least, because of the thermal inertia of ice-rich materials and where there are considerable thicknesses of permafrost. Thie (1974) reported a decrease in the areal extent of permafrost from 60 to 15% over the past several centuries in the discontinuous permafrost region of Manitoba. Suslov (1961) wrote that the permafrost at Mezen (NE of Archangel) has retreated northward at an average rate of 0.25 miles per year since 1837, and according to Bird (1967), Sumgin (1934) wrote that in the USSR generally, the southern limit of permafrost was probably receding, as a result of climatic amelioration.

In the zone of continuous permafrost, ground temperatures, which may be many degrees below 0C, would rise with a climatic warming, but perhaps no major changes in regional distribution of permafrost would occur. Meanwhile, however, we could expect progressive deepening of the active layer with the melting of shallow ground ice and ensuing thaw subsidence. In some areas, this would undoubtedly create severe maintenance and repair problems for roads, airports, buildings, pipelines, etc. Greater depths of gravel padding would be needed to preserve permafrost under roads and other structures.

In addition to permafrost degradation *per se*, *any* change in the temperature would cause major changes in the strength and deformation properties of frozen ground, even without thaw. (The effect of temperature is important not only because of its influence on the deformational mechanisms in the ice, but also as it determines the amount of unfrozen water in the frozen soil.) This could result in problems with the bearing capacity of piles, which are widely used in northern construction, as permafrost warms and adhesion forces decline. Other forms of foundations may also be affected. As well, the creep rates of (ice-rich) permafrost slopes would increase and slope stability would be decreased.

It is important to realise that climatic changes would also affect various earth surface processes which are characteristic of the permafrost environment, such as ice wedge cracking, frost heave, mass movements and creep. This may not be easy to determine, however, since according to Washburn (1980) the climatic relationships of permafrost processes are poorly known. Nonetheless, it has been hypothesised that thaw lakes in the western Arctic expanded during periods of warmer climatic conditions (e.g. Carson 1968). However, the sensitivity of thaw lakes to climate (change) is not well documented, although careful observations of lakes over a 10- to 20-year period might yield valuable information on growth patterns as a function of climatic variation (Burn and Smith, this conference).

Changes in precipitation

Consideration of climatic change should not be confined to temperature alone, nor simply to a change in mean annual conditions. According to Hansen et al. (1984), while the mean annual temperature in northwestern North America might be 7C higher in a 2xCO_2 world, increases in winter temperatures (up to 11C) would be 3 or 4-times greater than for summer (2 to 3C). In addition, there will be changes in precipitation (10 to 50% higher in summer, 60% higher in winter). Various earth surface processes in the permafrost environment could be affected by increased rainfall, but more importantly, perhaps, changes in snow cover would complicate the effect of climatic warming on ground thermal conditions.

While increased snow depths would partly offset any adjustment in ground thermal conditions to higher winter air temperatures, calculations made by Goodrich (1982) showed that a doubling of snow cover *per se* from 25 to 50 cm increased the minimum ground surface temperature by about 7C, and the mean annual surface temperature by 3.5C. If the 50 cm of snow builds up within 30 days in autumn, the minimum temperature would only reach -2.8C, the mean temperature would be +1.1C, and permafrost would degrade. In terms of ground temperatures, variations in snow cover are most critical at shallow depths, and the precipitation increases of as much as 60% in fall and early winter, predicted in the Hanser model, could be a significant factor in permafrost degradation, particularly in marginal areas.

When changes in precipitation are combined with the seasonal nature of predicted temperature increases, the problem of assessing climatic impacts on permafrost is made more complicated.

Significance of local conditions

Further caution must be exercised in simply extrapolating a warming trend in the atmosphere to the ground. Permafrost conditions are affected by the nature of vegetation, soil and snow conditions, as well as climate. Luthin and Guymon (1974) visualised these boundary layer interactions in terms of a buffer layer model, comprising the vegetation canopy, ground cover and snow cover, interposed between the atmosphere and the ground (Figure 1).

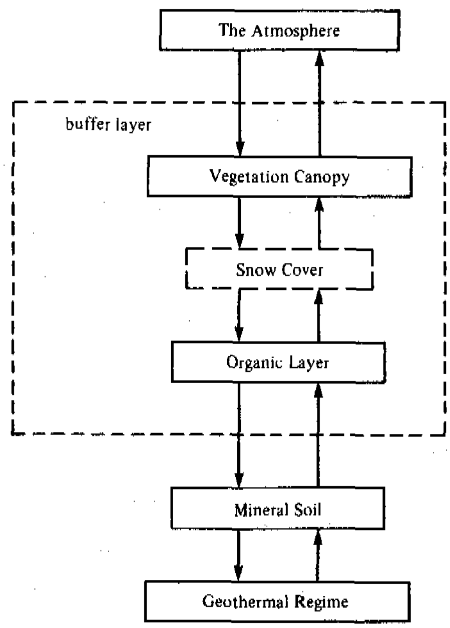

Figure 1. Boundary layer interactions affecting the ground thermal regime (after Luthin and Guymon 1974).

Atmospheric mass and energy flows together with the geothermal heat flux constitute the boundary conditions, with the vegetation canopy, snow cover (when present) and surface organic layer acting as buffers between the atmosphere and the ground. Consequently, wide variations in ground thermal conditions can occur within small areas of uniform climate due to local factors (e.g. Brown 1973, Smith 1975).

Figure 2, for example, shows there are significant differences in the annual ground temperature regime amongst neighbouring sites in the boreal forest region near Mayo, Yukon, Canada; the sites differ in terms of vegetation, ground cover and snow cover (the details of which may be omitted for the present purposes). The summer maximum 50-cm temperature varies from 0.5 (at site 6) to 7.1C (at site 4), and the winter minimum temperature from -2.9 (at sites 2 and 5) to -8.6C (at site 4). The fluctuations evident at 50 cm are highly damped by the 3-m depth. The mean annual temperature at 3 m varies from -0.1 at site 2 to -1.9C at site 6, where there is a 25-cm surface organic layer.

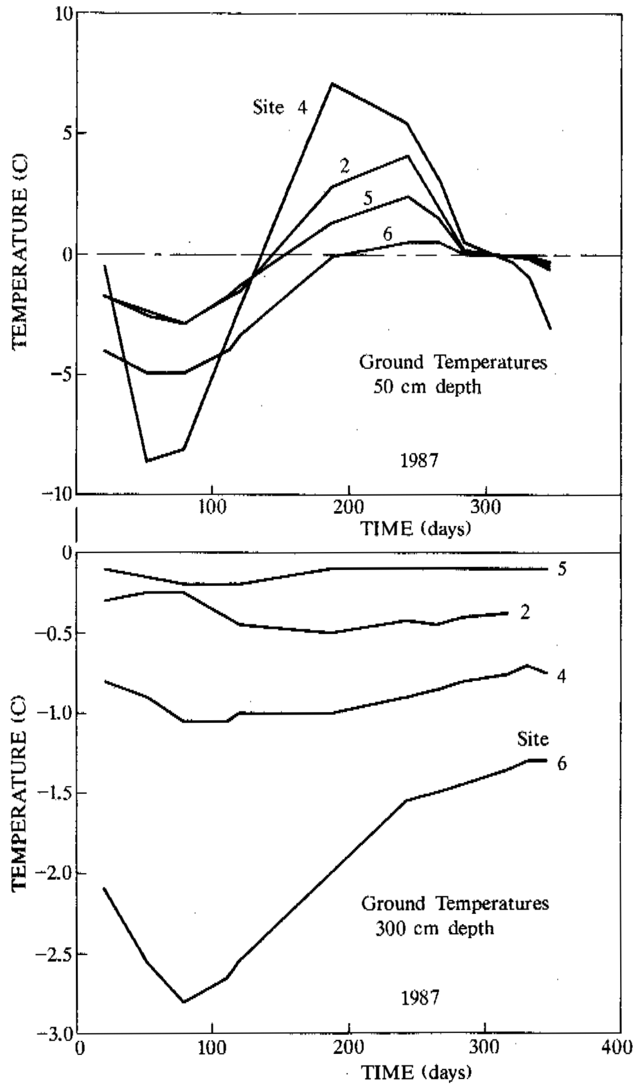

Figure 2. Annual ground temperatures at neighbouring sites near Mayo, Yukon, Canada.

The presence of organic material has been credited with the existence of permafrost in the southern margins (e.g. Lindsay and Odynsky 1965, Zoltai 1971), even where the mean annual surface temperature is above 0C (see Goodrich 1978). This implies that in similar areas further north, permafrost would not degrade nearly as quickly as at other sites with a climatic warming. Riseborough (1985) concluded that the ground thermal regime in the boreal forest, where a surface organic layer is present, will be considerably buffered from the effects of climatic change. The interception effects of the forest canopy, but, more importantly, the thermal resistance of the moss and peat, serve to isolate the ground from the atmosphere.

The fact that present ground thermal conditions, as illustrated in Figure 2, are so variably linked to the present climate makes it inevitable that they will be differentially buffered from the effects of changes in climate. In some cases, the response in permafrost conditions may be very slow to develop. In addition, ground thermal conditions will certainly be further affected as vegetation itself changes in response to climatic change. However, in areas of little vegetation or snowcover - such as can be found in the Canadian Arctic Islands, for example - the linkage between climate (air temperatures) and ground temperatures is more direct, and so should be the effects of climate change.

Importance of lithologic conditions

The thermal stability of permafrost will be aided by the occurrence of ice-rich material near the present permafrost table. Figure 3 shows the results of a simple simulation of ground temperature changes (using the model of Birch, 1948), for an increase in the surface temperature of 1C per decade over a sixty-year period. Mean annual temperature profiles at the end of the period are shown for three values of the ground thermal diffusivity. (An initial equilibrium profile is included for reference.)

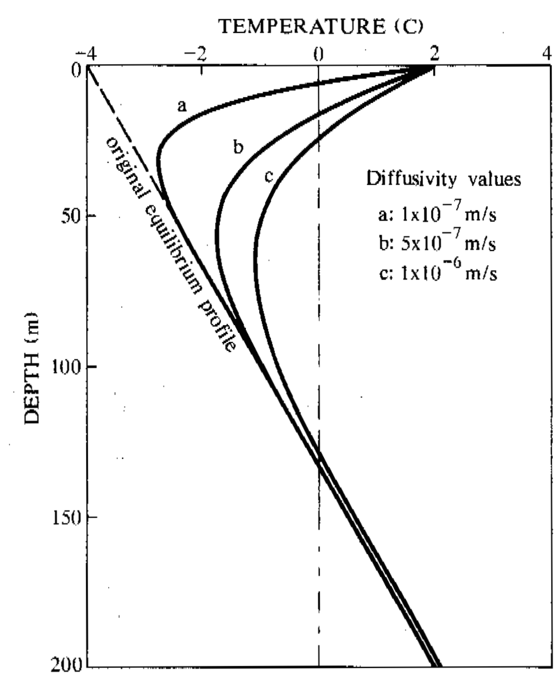

Figure 3. Estimated ground temperature profiles following a climatic warming.

The latent heat effects of ground thawing must be considered in this problem, since they will largely determine the rate of permafrost degradation. By using apparent diffusivities in the model (e.g. see Smith and Riseborough 1985), the effects of phase transition can be mimicked. A diffusivity value of $1x10^{-7}$ m^2 s^{-1} repesents the thawing of icy permafrost, while the value of $1x10^{-6}$ applies better to rock or unsaturated frozen (sandy) soil. There are striking differences in the depth of permafrost degradation, which ranges from only 6 m in the ice-rich case to 26 m for the highest diffusivity. In the latter case, the base of permafrost also rises with the general warming in the profile, but only by about 2 m. After 500 years, the permafrost table in the icy case would still have degraded to only about 40 m, and the base of permafrost would have risen by 4 or 5 m.

Consideration of the local surface and subsurface conditions indicates the difficulty of assessing the sensitivity of localised permafrost conditions to regional climatic change (see also, Smith and Riseborough 1983).

PROCESSES AND FEATURES OF PERMAFROST ENVIRONMENTS

Slope movements

Typical features of slope failure and related phenomena of mass-wasting in the permafrost environment are described in the literature (e.g. French 1976). Most of the problems arise as a result of thawing with consequent loss of strength - e.g.

-mudflows result from the thawing of ice-rich permafrost, and flow- dominated failures on natural slopes and cut slopes are common in permafrost areas;

-skin flows (active layer glides) involve the detachment of a thin veneer of vegetation and mineral soil with subsequent movement over the permafrost table. These flows are commonly found in ice-rich sediments, and result from the thaw failure of the ice-rich contact zone. Any surface disturbance (e.g.forest fire) or climatic change which increases the thaw depth will promote such failures (e.g. Mackay and Mathews 1973);

-slumping, although widespread in many regions, is especially common in the permafrost environment, occurring wherever ice-rich soils are subject to thawing;

-solifluction is also prominent in the permafrost environment, because the permafrost table prevents downward movement of water leading to conditions of saturation, and also acts as a shear plane for failure.

While there are few studies which links mass movements to climatic conditions (e.g. Lewkowicz 1987), it seems reasonable to suppose that some increase would occur as a result of changes in climate and degradation of permafrost. Deepening of the active layer (with consequent melting of ice-rich soil) could lead to conditions of instability and failure, as also could increased summer rainfall (cf.Mackay and Mathews 1973). In addition to the displacements of thawed ground, there is a need to consider the possibility for increased deformation of frozen ground under conditions of warming.

Strength and creep of frozen ground

The mechanical behaviour of frozen soils is very different from that of unfrozen soil, because of the presence of ice and unfrozen water. Frozen soils are more subject to creep and relaxation effects, and their behaviour is strongly affected by temperature change. There is extremely limited information on the creep of natural permafrost (or ice) in the range of stress and temperature (i.e. near 0C) of practical interest (Morgenstern 1985). If we assume that the creep deformation of ice-rich permafrost will be dominated by the rheological properties of the ice, then we can look to Glen's flow law for ice. Based on this, at a temperature of -22C the strain rate produced by a given stress is only one-tenth of its value at 0C, and the strain rate almost doubles from -5C to 0C. Moreover, we should remember that a fine-grained soil close to 0C will contain a considerable amount of unfrozen water, and thus the viscosity will be less than for pure ice. This implies faster deformation rates, and it is by no means clear, therefore, that the flow law for ice will provide us with conservative estimates of strain rates.

Ice wedge cracking

Washburn (1980) states that while the relation of frost cracking to (mean) air or ground temperatures is of great interest, the relationship is only approximately known. Based upon the distribution of active ice wedges in Alaska, Pewe (1966) suggested that ice wedges only form where the mean annual temperature is -6C or colder, although Mackay and Mathews (1973) reported active ice wedges near Fort Good Hope in the discontinuous zone. In any event, authors seem to agree that a rapid drop in the ground temperature is necessary to cause cracking - this would certainly be impeded by the insulating properties of a deep snow cover. For example, Mackay (1984) has shown that a snow depth of 60cm was sufficient to prevent ice-wedge cracking in an area of active ice wedges on Garry Island. (See his figure 5, of minimum ground temperature vs. snow depth).

Other features

Tarnocai and Zoltai (1978) reported on the distribution of earth hummocks in the western Canadian Arctic. They concluded that the hummocks developed only during the cool climatic period since 5000 B.P.; the most southerlv hummocks (e.g. Fort Simpson, South Indian Lake) must have developed during the coldest sub-period of the last 5000 years, and became inactive when climatic conditions moderated (i.e.cryoturbation apparently ceased a long time ago). Zoltai (1975) analysed the pattern of reaction wood in tree ring records, to infer the incidence of ground instability in earth hummocks at Inuvik. The record indicates a period of greater ground instability (heaving?) between about 1850 and 1950, with reduced activity since 1950. However, if this pattern is related to the climatic warming between 1850-1950 and the cooling trend since, it would seem to contradict the conclusions of Tarnocai and Zoltai (1978).

SUMMARY

The effects of climatic change on the permafrost environment should be considered in a variety of ways:

1. There would be long-term changes in the distribution of permafrost, as it degrades under a warmer climate. Tens of thousands of square kilometers of permafrost which are warmer than -3C would disappear, eventually, under the predicted climatic warming. In some areas, however, complete degradation would take many centuries, at least, because of the thermal inertia of ice-rich materials and considerable thicknesses of permafrost.

2. More importantly, perhaps, there would be various rapid onset effects, especially in the discontinuous zone, where most permafrost is either relict, or in such delicate balance that climatic or other environmental changes could have drastic disequilibrium effects. We should expect progressive deepening of the active layer with the melting of shallow ground ice, leading to widespread thermokarst and increased slope instability. In some areas, this would undoubtedly create severe maintenance and repair problems for all range of structures.

3. Even without thaw, change in permafrost temperatures would cause (major) changes in the strength and deformation properties of the frozen ground. This could lead to problems with the bearing capacity of piles, which are widely used in northern construction, as permafrost warms and adhesion forces decline. Also, the creep rates of (ice-rich) permafrost slopes would increase and slope stability would be decreased.

4. The rate of various permafrost processes, such as ice wedge cracking, frost heave, mass movements (and creep), for example, would change. However, the climatic relationships of permafrost processes are poorly known.

The consideration of climatic change should not be confined to temperature alone, nor simply to a change in mean annual conditions. According to Hansen et al. (1984), increases in winter temperatures would be 3 or 4-times greater than for summer. In addition, there will be increases in precipitation in both winter and summer. Various earth surface processes in the permafrost environment could be affected by increased rainfall, but more importantly, perhaps, changes in snow cover would complicate the effect of climatic warming on ground thermal conditions.

Further, we must appreciate that more than simply climatic conditions determine the surface temperature regime and thermal conditions in the ground. While the occurrence of permafrost depends in a broad way on climate, the effects of climate and climatic change may be modulated, perhaps strongly, by local microclimatic (ecological) and lithologic conditions.

With a climatic warming it is inevitable that some permafrost would eventually disappear and ultimately there would be a new "permafrost map". However, the more immediate question is how quickly the effects will be seen, and at what rate will they develop? And, would such changes be important to northern geotechnical engineering? Since this concerns the <u>transient</u> aspects of the problem (rather than the eventual steady-state), some important questions are:

1) What is the rate and magnitude of permafrost change due to climatic warming likely to be? What is the thermal inertia (time constant) of permafrost sediments?

2) How can we assess the effects of regional climatic change on permafrost conditions, given that local conditions exert a strong influence on the ground thermal regime? i.e. How are ground thermal conditions buffered from (or linked to) changes in climate (including feedback effects)?

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CLIMATIC CONDITIONS AND PERMAFROST DEVELOPMENT ON THE SVALBARD ARCHIPELAGO

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Abstract:

The climate on the Svalbard Archipelago is described using synopsis measurements from several meteorological stations. Thawing and freezing indexes for Longyearbyen, Svea, Ny-Ålesund, Hopen, Isfjord and Bjørnøya are calculated based on the calculation of a 'synthetic' year for the weather stations. The period between 1980 and 1987 has been chosen so as to enable comparison between the stations. For these years the climate has been markedly cooler than in the periods 1920 to 1960 and 1971 to 1977. The average annual temperature is generally lower on Svalbard by approximately 2°C, when compared to the average temperatures in 1920 to 1960. Should this cooler weather last for a significant period of time, it is expected that the average permafrost depth on Svalbard will increase with approximately 40 to 80 m, depending on local variations in climate and ground characteristics.

The climate of the Svalbard Archipelago

The climate of the Svalbard Archipelago has been described in terms of a statistical treatment of weather synopsis data at several locations on the island group. (Steffensen, Esther, 1969 and 1982). Synopsis data exist from as early as 1910. Even older data have been recorded, often by hunters or by whaling and sealing ships, but these records have not been transferred to computer files and are not as yet available for statistical analysis. Thus, the data used in describing the climatic conditions on Svalbard are mainly from the period 1957-1987.

Conditions peculiar to the Svalbard archipelago are the marked annual variation in radiation as compared to diurnal; this is particularly noticable during the 3-4 summer months with midnight sun and net radiation heat gain, and in the 3-4 winter months with the sun below the horison and net radiation heat loss. The large scale atmospheric circulation is characterised by the influence of the mild air originating from the low pressure area around Iceland, and colder air from Greenland and the Arctic Ocean. Ocean currents also influence the weather on Svalbard. The warm Norwegian current flows partly into the Barents Sea and partly towards the west coast of Spitsbergen, while a cold southwest-bound current moves towards Bjørnøya. The competing air masses and currents give rise to widely varying weather conditions, particularly on the west coast of Spitsbergen.

Steffensen (1982) has classified the climate at the Svalbard stations as either Tundra Climate (with the mean monthly air temperature above 0°C at least one month during the year) or Polar Marine with a marked marine influence on weather patterns. Sometimes the weather at a single station can vary between the two classifications. Examples are Jan Mayen and Bjørnøya.

The most remarkable features of air temperature measurements at Svalbard, are the rapid and significant air temperature fluctuations, particularly during winter. These temperature fluctuations do not influence the developement of permafrost depth to any large extent, but large variations in the summer temperatures may have a significant effect on the depth of the upper, active layer of permafrost.

The calculation of 'synthetic' years.

As a tool for calculating mean freezing and thawing parameters at locations of the meteorological stations at Svalbard, mean temperatures of each day in the year for a spesified period of time have been calculated. These temperature means form a 'synthetic' year which characterise the temperature variations dependent on location (S. Lystad, pers. com. 1988). The small-scale fluctuations which are characteristic of the winter months (January and February) are evened out in the 'synthetic' representations, but local differences in the temperature fluctuations are preserved. An example of a 'syntetic' year for Svea is given in table I.

Table I :

Day	Jan	Feb	Mar	Apr	May	Jun
1	-12.8	-14.9	-15.3	-14.0	-8.1	0.5
2	-13.7	-16.2	-16.2	-12.6	-7.9	0.5
3	-14.3	-15.0	-17.7	-14.0	-7.7	-0.2
4	-15.5	-16.1	-19.6	-12.3	-7.3	-0.6
5	-15.4	-17.1	-16.7	-11.3	-7.8	0.6

Day	Jan	Feb	Mar	Apr	May	Jun
6	-15.7	-18.4	-12.8	-12.7	-7.6	0.6
7	-15.8	-16.7	-9.9	-14.5	-7.3	0.3
8	-14.1	-12.6	-9.7	-16.2	-6.6	0.5
9	-12.8	-10.7	-8.1	-15.6	-6.3	1.0
10	-14.3	-13.7	-6.8	-16.2	-6.2	1.1
11	-12.1	-14.9	-5.9	-16.7	-4.5	1.6
12	-11.1	-15.5	-8.9	-16.1	-5.2	1.8
13	-13.3	-15.0	-10.9	-14.2	-4.5	1.8
14	-14.8	-16.0	-11.1	-13.6	-4.5	2.0
15	-17.6	-17.7	-14.5	-13.1	-3.4	2.2
16	-19.2	-20.3	-14.5	-12.9	-2.0	2.3
17	-17.4	-17.3	-13.2	-11.8	-2.3	2.3
18	-17.1	-13.2	-12.3	-12.8	-2.2	2.7
19	-16.6	-14.9	-17.4	-13.1	-2.0	2.6
20	-18.0	-15.4	-12.6	-9.3	-1.4	2.9
21	-19.0	-12.2	-12.9	-8.6	-1.5	3.1
22	-16.9	-15.2	-12.6	-7.7	-1.6	3.4
23	-15.9	-16.1	-14.6	-8.7	-2.2	3.9
24	-15.5	-15.0	-14.8	-9.7	-2.8	4.3
25	-14.5	-17.1	-16.5	-10.0	-2.9	4.4
26	-17.0	-17.5	-17.4	-8.7	-1.9	3.9
27	-18.8	-17.0	-19.6	-9.4	-1.5	4.5
28	-20.0	-14.6	-18.4	-8.6	-1.6	4.8
29	-18.7		-17.8	-8.8	-1.5	5.3
30	-18.5	-	-18.6	-9.0	-1.1	6.6
31	-16.8	-	-17.5	-	-0.1	-

Day	Jul	Aug	Sep	Okt	Nov	Des
1	6.1	6.3	2.7	-1.8	-9.0	-13.0
2	5.4	6.9	2.5	-2.6	-9.6	-12.9
3	5.6	7.0	3.0	-3.4	-9.5	-14.3
4	5.6	6.0	2.3	-2.6	-10.9	-17.7
6	5.5	5.8	2.4	-3.6	-11.5	-16.8
7	6.0	6.0	2.0	-3.1	-10.3	-16.5
8	6.1	5.5	2.0	-2.5	-11.3	-16.9
9	5.6	5.5	1.9	-4.0	-11.6	-18.1
10	5.8	5.3	1.6	-3.8	-12.6	-20.4
11	6.6	5.6	1.9	-3.7	-12.7	-18.8
12	6.0	6.0	1.5	-3.9	-9.5	-17.0
13	5.5	6.3	0.7	-4.6	-10.1	-14.3
14	5.4	5.7	0.5	-4.9	-10.8	-16.0
15	5.6	4.7	1.0	-4.2	-10.8	-17.8
16	6.1	5.0	0.7	-5.0	-9.2	-18.9
17	6.0	4.8	-0.3	-6.8	-9.3	-17.0
18	6.2	4.2	0.2	-6.8	-10.1	-15.3
19	6.8	4.3	0.6	-7.1	-10.4	-15.9
20	6.2	4.4	-0.6	-7.7	-9.5	-17.5
21	5.6	3.4	-1.8	-6.9	-12.1	-13.0
23	6.4	3.5	-1.4	-6.2	-12.3	-12.3
24	6.2	3.5	-1.1	-5.6	-13.2	-11.9
25	6.2	3.6	-2.1	-7.1	-11.6	-11.4
26	6.9	3.5	-2.7	-7.3	-12.6	-14.4
27	6.8	3.0	-3.2	-7.0	-14.7	-16.7
28	6.4	3.5	-2.8	-9.1	-15.1	-15.4
29	6.0	3.3	-2.7	-9.4	-11.8	-13.9
30	6.1	3.2	-2.4	-9.2	-13.2	-13.6
31	6.5	3.0	-	-9.2	-	-13.2

The different meteorological stations on Svalbard have recorded data over different time periods. Thus, to enable comparison of permafrost conditions between the stations, the period between 1980 and 1987 has been chosen as the basis for calculations of ´synthetic´ years. The exception is for Isfjord Radio where synopsis observations stopped in 1975. For Isfjord Radio a ´synthetic´ year for the period 1961 to 1968 was calculated. The mean daily temperature is calculated for all the stations, but it is feasible with this method to calculate ´synthetic´ years for all the synopsis data, i.e. measurements at 0700, 1300 and 1900 hours, and for all weather parameters. Calculation of freezing and thawing indexes have been made from mean daily temperatures.

Permafrost developement.

No attempt has been made to use emphirical data to calculate thawing and freezing indexes of the ground. This is because too little is known about the soil and other local geographical features at the weather stations, which are significant to permafrost developement. However, other significant climatic parameters are given for each station and presented in table II.

Table II:

Station	Temp. annual mean	Clouds annual mean	No. of clear days	Prec.* no of days	Thawing index (air)	Freezing index (air)
LYB	-6.0	5.3/8	54	93	10095	62232
NÅ	-5.8	5.4/8	52	100	8088	58728
S	-6.7	-	-	-	9907	68064
IS	-5.9	5.6/8	34	116	7356	58752
BJ	-2.0	6.5/8	7	153	9960	26993
HO	-6.2	6.0/8	27	171	4176	57600

LYB=Longyearbyen, Svalbard Airport, Posn.: 78°15´N, 15°28´E
NÅ=Ny-Ålesund, Posn.: 78°55´N, 11°56´E
S=Svea, Posn.: 77°54´N, 16°41´E
IS=Isfjord Radio, Posn.: 78°04´N, 13°38´E
BJ=Bjørnøya, Posn.: 74°31´N, 19°01´E
HO=Hopen, Posn.: 76°30´N, 25°04´E

* Numbers given in this row refer to days with snow fall more than 0.1 mm. Since measurements of snow depth are not consistent, these cannot be statistically treated. The number of days with snow fall more than 0.1 mm (water equivalent) gives an indication of snow depth, but unfortunately snow cover duration is not reported consistently.

The thawing index may be used as an indication of the mean depth of the active layer. Comparing the different weather stations, Longyearbyen, Svea and Bjørnøya all have large thawing indexes. The thawing index of soil as compared to that of air is larger if the weather is clear. From table II it is seen that Bjørnøya have very few clear days on average, but this may be counteracted by the fact that there is much precipitation, and particularly rainfall during summer at Bjørnøya. The thawing index of Hopen is very low, and the number of clear days is also low. It may be assumed that the active layer on Hopen is more shallow than in Longyearbyen. It is not possible to be more presise based on meteorological data alone.

The freezing index together with the mean annual air temperature can be indicative of the permafrost depth at the weather stations. Due to its southerly position and marine climate, Bjørnøya has a much higher mean annual air temperature and a lower freezing index than the other stations. It is therefore likely that the permafrost layer at Bjørnøya is no deeper than approximately 80 m. During the warm climatic period between 1920 and 1960, the permafrost layer on this island may have disappeared altogether.

Svea and Longyearbyen have the largest freezing indexes and the lowest mean air temperatures. It is therefore likely that the permafrost is deepest at these two stations. Again, the results are likely to be modified by local geographical and soil conditions.

Conclusions

A condition for the existence of permanentely frozen ground, is that the mean air temperature is below 0°C. Permafrost depth and cover is dependent on many features of the climate and the landscape, such as variations in snow cover, the presence of melt water in the summer, the type of ground and the top vegetation, as well as the orientation of slopes.
At various locations on the Svalbard Archipelago, permafrost depths have been measured between 75 and 450 m (Liestøl, 1980).

Permafrost is maintained by mean annual subzero temperatures, while the geothermal heat radiation thaws the ground from beneath. The gradient of the geothermal heat radiation has been measured in several boreholes to be approximately 1°C per 40m (Liestøl, 1980). From the local mean annual air temperature, it is therefore possible to estimate the depth of permafrost. Approximate permafrost depths at different meteorological stations based on annual mean air temperature are given in table III.

Table III:

Station	Permafrost depth (m)	Active layer depths (m)
LYB	240	0.9
NÅ	232	0.8
S	268	0.9
IS	236	0.8
BJ	80	1.1
HO	248	0.7

Climatic changes in air temperatures can be detected as fluctuations in running five-year mean temperatures. Steffensen (1982) has shown that the Svalbard Archipelago has experienced two periods of warmer weather in recent times; from approximately 1920 to 1960, and between approximately 1971 to 1977. In both these periods the mean annual air temperature was approximately 2°C higher than today, although there are variations between the different meteorological stations.

Measurements made by H. Major in 1956 (Liestøl, 1980) in boreholes near Svea and Longyearbyen, confirm the depth of permafrost estimated by mean annual air temperaure, and also show a characteristic 'bend' in the upper parts of the temperature curves which corresponds to the warmer climate in 1920 to1960. The propagation speed of the 'heat' wave as estimated from these boreholes, is approximately 3 m per year. It is likely that temperature changes will 'propagate' in the soil at different speeds depending on local characteristics.

However, as a very rough estimate (which should be regarded as an indication only) it is possible to calculate the depth of the active layer by multiplying an average propagation of melting (3 m per year) with the number of days when the air temperature is above zero. No corrections have been made to accomodate the fact that the soil temperature may be significantly higher than the air temperature under certain weather conditions. Melt water also has a significant influence on the depth of the active layer. The active layer depths presented in table III should therefore be regarded as low estimates.

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PERMAFROST AGGRADATION AND DEGRADATION ON ARCTIC COASTS OF NORTH AMERICA

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SYNOPSIS Permafrost conditions on the arctic coasts of North America are related to the Pleistocene history of the area and the relative changes in sea level. In general, the western Arctic is presently experiencing rising sea levels with submergence, coastal recession and permafrost degradation; whereas the eastern Arctic and the Arctic Archipelago are experiencing isostatic uplift and permafrost aggradation at emergent shorelines. The effect of these changes on ground temperatures, and on the occurrence of ice-bonding, depends on such factors as sea water temperatures and salinities as well as local thermal effects.

INTRODUCTION

Research on permafrost in the coastal zone of arctic North America has increased substantially over the last decade as a result of accelerated economic development of the north. The main incentive for such interest has been the discovery of significant petroleum reserves beneath the Beaufort Sea continental shelf and in the Arctic Archipelago. Increased industrial activity and the eventual transport of hydrocarbons to southern markets will require even more detailed information on permafrost conditions for the planning and construction of port facilities, pipeline landfalls and other nearshore and shoreline facilities. Both government regional surveys and industrial site-specific projects have contributed to an increased knowledge of coastal permafrost conditions through the study of coastal processes and through geotechnical evaluations. However, most work has been restricted to areas of economic interest, and large tracts of coastline have yet to be examined.

This paper attempts to characterize (in a general manner) the different temperature regimes of permafrost in the coastal areas of arctic North America, from studies of North American researchers. The role of permafrost in geomorphic processes in these areas is the subject of a companion paper by H.J. Walker (this volume).

GLACIAL HISTORY

The nature and distribution of permafrost along the present-day coastlines of arctic North America are, to a large extent, defined by the Pleistocene history of the area. Specifically, the permafrost characteristics depend on the local history of ground surface temperatures at any given site. Thick "cold" permafrost has formed in areas that have been aerially exposed for long periods of time; whereas in areas which have been covered by water (sea or lakes) or glacial ice, the growth of permafrost has been inhibited and any pre-existing permafrost may have degraded.

Extensive glaciations of arctic regions in Pleistocene times have strongly influenced the ground surface temperature history of many arctic coastline areas. The glaciations resulted in major changes in relative sea level, partly due to the transfer of water from the oceans to the glaciers, and partly due to the depression of the land by thick ice masses. The lower sea levels exposed the continental shelf areas of the Beaufort and Chukchi Seas, and as a result, permafrost formed in these areas. In contrast, during the glacial periods the Arctic Islands were largely ice-covered and the land areas were depressed; they are presently experiencing post-glacial isostatic uplift. Thus, the glacial history of arctic North America has contributed to the existence of two zones with distinctly different coastal characteristics. That is, at the present time, coastal recession and erosion in the western Arctic result in the degradation of permafrost, while land is emerging and permafrost is aggrading in the eastern Arctic (Figure 1).

The extent of Pleistocene glaciations in the North American Arctic are summarized in Figure 1 (after Prest, 1984). The late Wisconsinan glacial advance is the best-documented, but there is evidence that older Pleistocene glaciations covered the entire Arctic Achipelago and much of the Canadian Beaufort Sea coast and continental shelf. In Alaska, ice was restricted to a relatively small inland area during the Pleistocene glaciations and did not reach the Beaufort Sea coastline.

The late Wisconsinan glacial advance did not extend onto the Beaufort Sea continental shelf except in isolated areas. Figure 1 shows the large areas of the Beaufort and Chukchi continental shelves which would have been exposed at times of maximum sea level lowering. These areas were exposed as dry land, and may have been subjected to mean annual ground surface temperatures much lower than those of today (Brigham and Miller, 1983).

Estimates of the extent and duration of sea level lowering (Mackay, 1972; Hill et al., 1985) suggest that extensive permafrost would have developed beneath these present-day shelf areas. Permafrost thicknesses in excess of 700 m are known to exist beneath the Canadian Beaufort Sea shelf today (Mackay, 1986b), though it is possible that this thick layer may be a result of more than one period of permafrost aggradation brought about by a lowering in sea level.

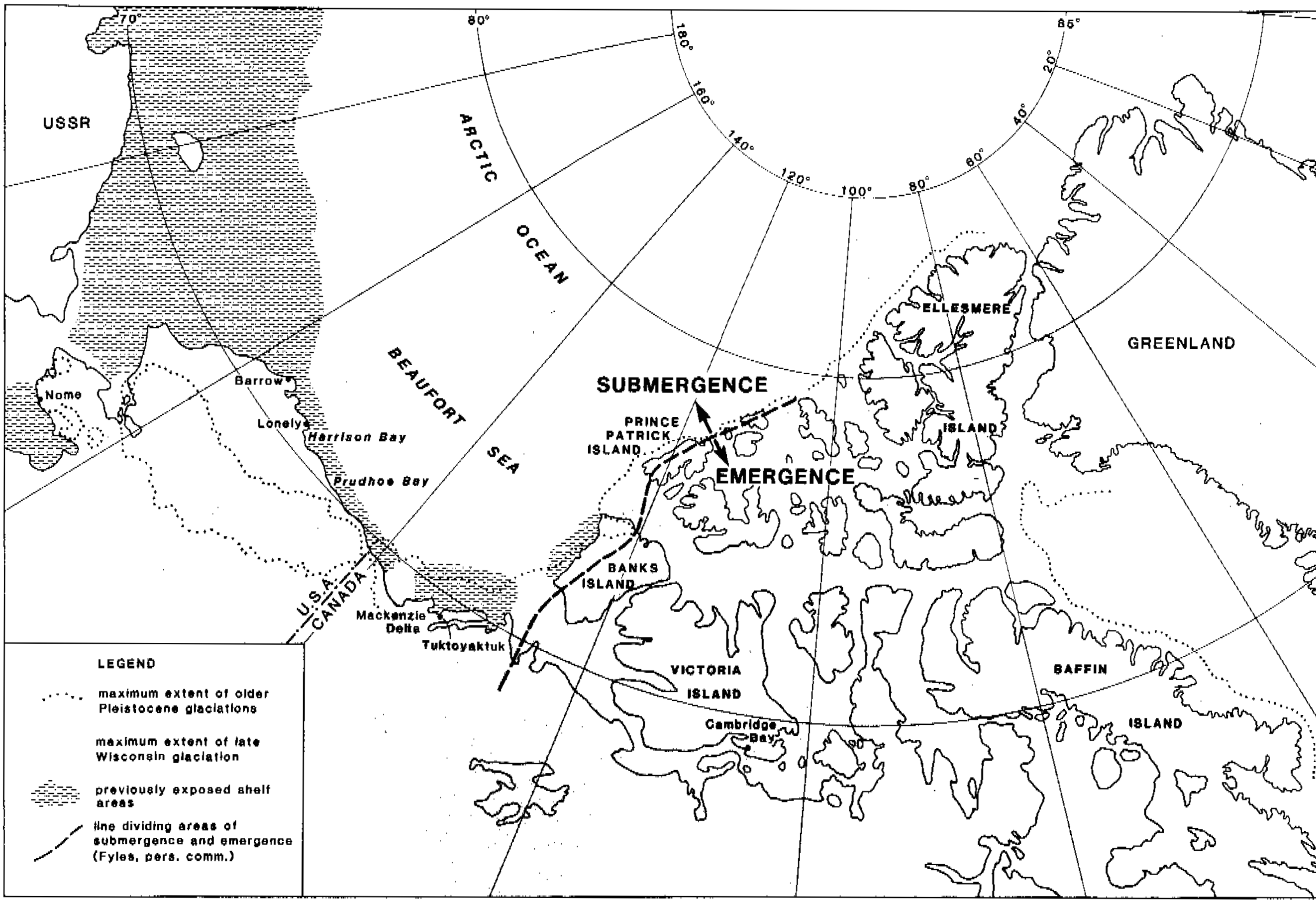

Figure 1 Map of the North American Arctic showing the maximum extent of major Pleistocene glaciations and previously exposed shelf areas (after Prest, 1984).

COASTAL RECESSION AND PERMAFROST DEGRADATION

Canadian Beaufort Sea

Evidence for coastal recession can be found along the length of the Beaufort Sea coast, where coastal retreat rates in places can exceed 15 m/year. Studies in the Mackenzie Delta region measured rates of between 1 and 13 m/year (Forbes and Frobel, 1985; Mackay, 1986b). Drowned coastlines also persist along the west coast of Banks Island and Prince Patrick Island (J.G. Fyles, pers. comm.). As the sea encroaches, degradation of permafrost occurs because the mean annual ground surface temperature can be no colder than seawater (-1.8°C). This is a slow process, however; sub-bottom temperatures as low as -4.2°C have been recorded beneath the Beaufort Sea shelf in water depths of 50 m (Weaver and Stewart, 1982; Mackay, 1986b).

In the Mackenzie Delta area of the Beaufort Sea, thermal degradation is accelerated by the seasonal inundation of large amounts of warm (>0°C) fresh water from the Mackenzie River. Evidence of submerged channels, some of which are associated with the configuration of the modern day Mackenzie River channels, can be seen in the bathymetry of the Beaufort shelf. This suggests that the influence of warm river water may have been a significant factor in permafrost degradation, at least in local regions, for the last several thousand years.

Under present day conditions, warm mixed (river and sea) water is resident on the seabottom during the summer months to water depths beyond 20 m, and distances up to 50 km offshore. In winter months the fresh water flow is restricted and cold seawater invades the inshore region. Figure 2 shows the extent of this seasonal zone in the Mackenzie Delta area. The active layer created by the seasonal bottom temperature variation is discussed in detail by Hunter et al. (1988). The net thermal effect is the development of a nearshore thaw zone in the seabottom. For water depths greater than 8-10 m, however, the mean annual bottom temperature is again below 0°C, and permafrost could be re-established below the active layer on the seafloor.

A diagrammatic model of the nearshore zone in the Mackenzie Delta area, illustrating the degradation of permafrost under the influence of warm river water, is shown in Figure 3. Thermal degradation of permafrost in the inshore regions (water depths <8-10 m), where mean annual seabottom temperatures are above 0°C, is well documented (Hunter et al., 1988), with thaw depths reaching 50 m or more. The bottom of this thaw zone is closely associated with the 0°C isotherm, and poorly bonded permafrost is evident (Dallimore et al., 1988).

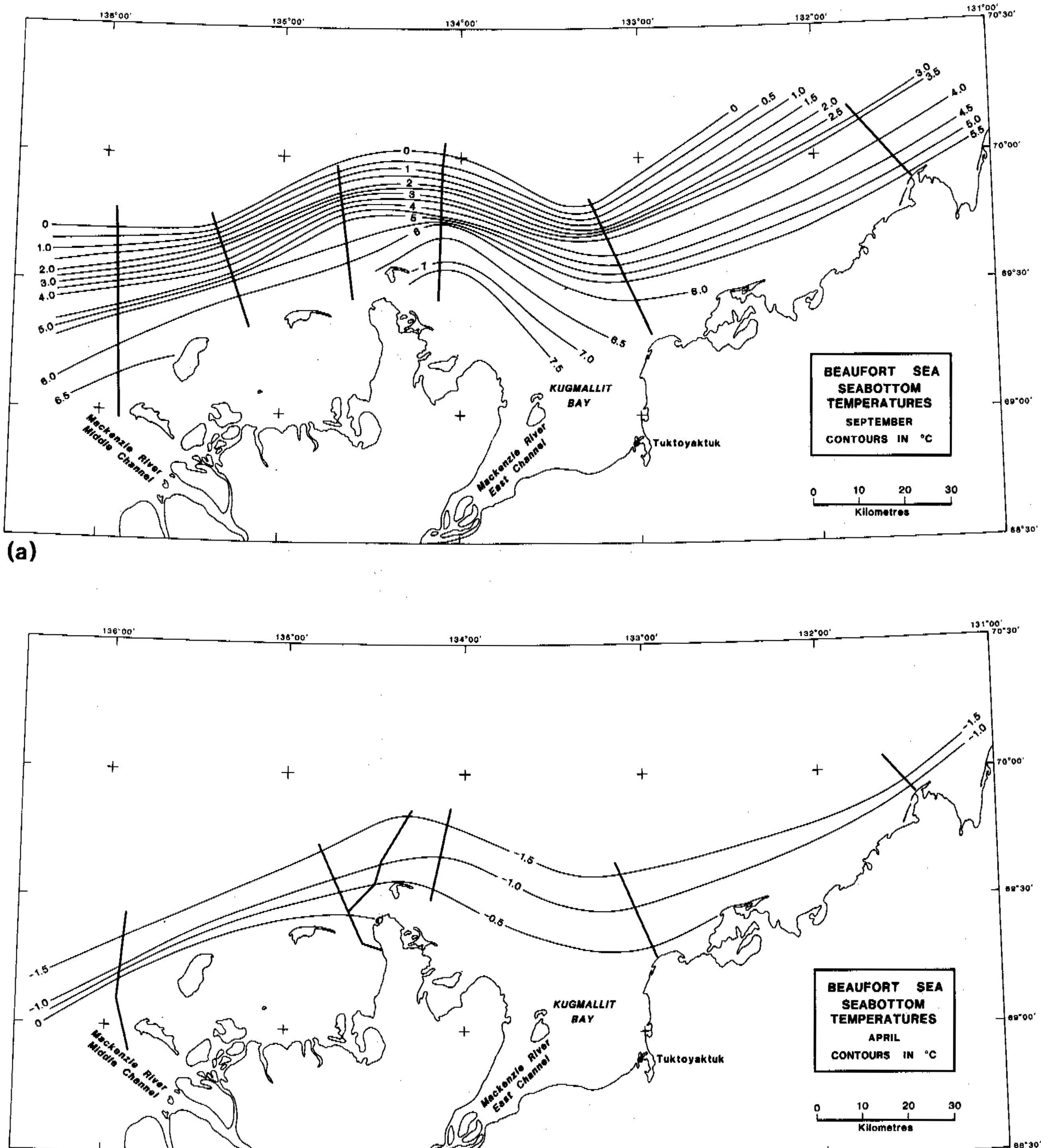

Figure 2 Areal distribution of seabottom temperatures for the southern Beaufort Sea from a) end-of-summer, and b) end-of-winter measurements. The heavy lines indicate the onshore-offshore transects along which the data were collected.

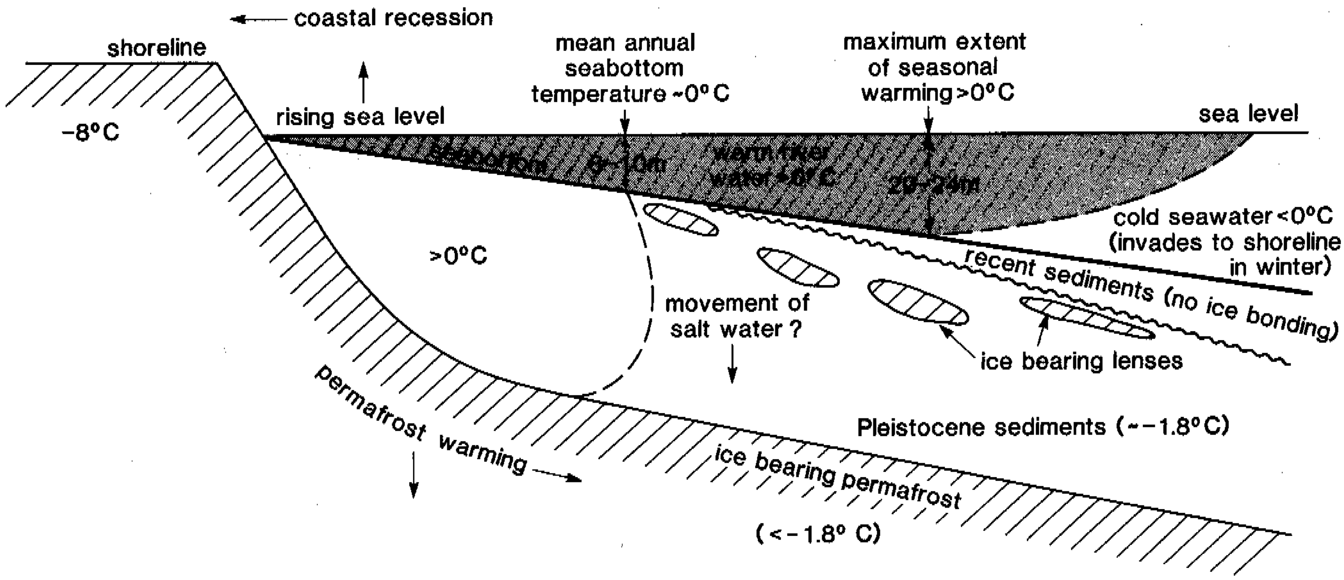

Figure 3 Permafrost conditions in the nearshore region of the Mackenzie Delta coast.

There is seismic evidence from some of the farther offshore areas of the Beaufort Sea shelf which suggests that this relict boundary persists at depth (Hunter, 1984), as indicated in Figure 3. No continuous ice-bonded permafrost occurs between seabottom and this boundary, even though sediment temperatures may be as low as -1.8°C. Because the temperature at the seafloor can be only marginally below 0°C (>-2°C), whether or not ice-bearing permafrost could be re-established in the sub-bottom sediments is critically dependent upon the original salinity of the pore water, the rate of salt transport into the bottom sediments after inundation, and on the grain size of the sediments. However, there is ample evidence of the occurrence of lenses of ice-bearing sediments within a few metres of the seafloor in water depths beyond 8 m (Blasco, 1983; Fortin et al., 1987). These may have refrozen in-situ with rising sea levels.

The deeper thermal effect of coastal recession in the Tuktoyaktuk Peninsula area of the Beaufort Sea is illustrated in Figure 4 (after Taylor and Judge, 1988). Temperature observations from a well situated 2 km offshore have been interpreted to represent submergence for a period of approximately 300 years, implying a coastal recession rate of 6-7 m/year. At depth, the negative temperature gradient is a result of the warming effect of the seawater superimposed on the pre-submergence temperature gradient. Analysis suggests that the mean annual surface temperature prior to submergence was -12°C; somewhat colder than present day values.

In the immediate shoreline area (water depths of <2 m) the temperature regime of the shallow permafrost is dominated by the insulating effect of snow cover, and the cooling effect of sea-ice frozen to bottom. This is illustrated in a diagrammatic model of the shoreline zone shown in Figure 5, along with several sub-bottom temperature profiles from an onshore-offshore transect in the Mackenzie Delta area.

The thickness of snow cover is governed by the morphology of the coastline and its orientation with respect to the prevailing winds. In the coastal zone of the Beaufort Sea, shoreline scarps of 2 m or more in height are common, and act as traps for snow accumulation during winter. Such a thermal blanket results in warmer shallow permafrost temperatures beneath the narrow beach zone than beneath the cliff-top or immediately offshore (Fig. 5).

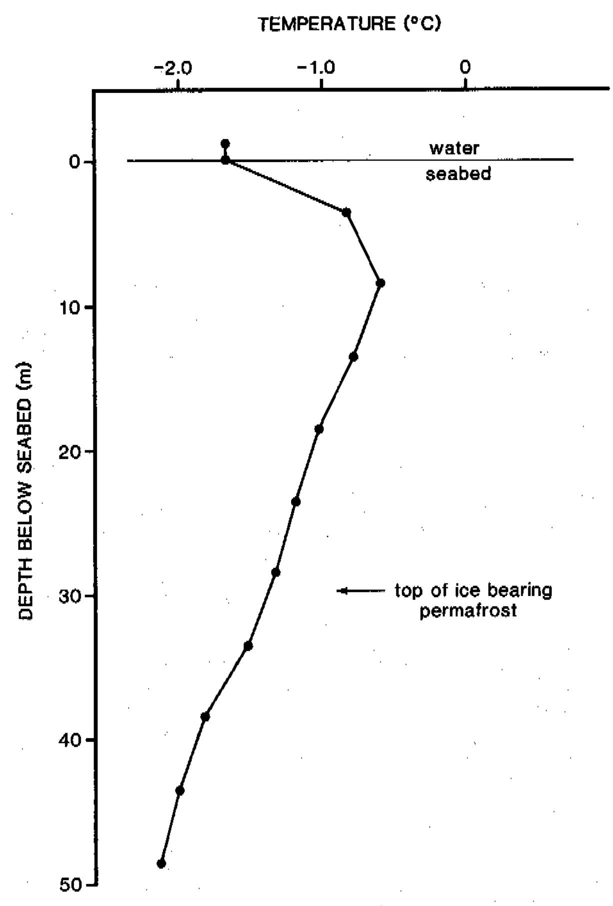

Figure 4 Temperature-depth profile measured in a geotechnical hole drilled 2 km offshore in the eastern Beaufort Sea (from Taylor and Judge, 1988).

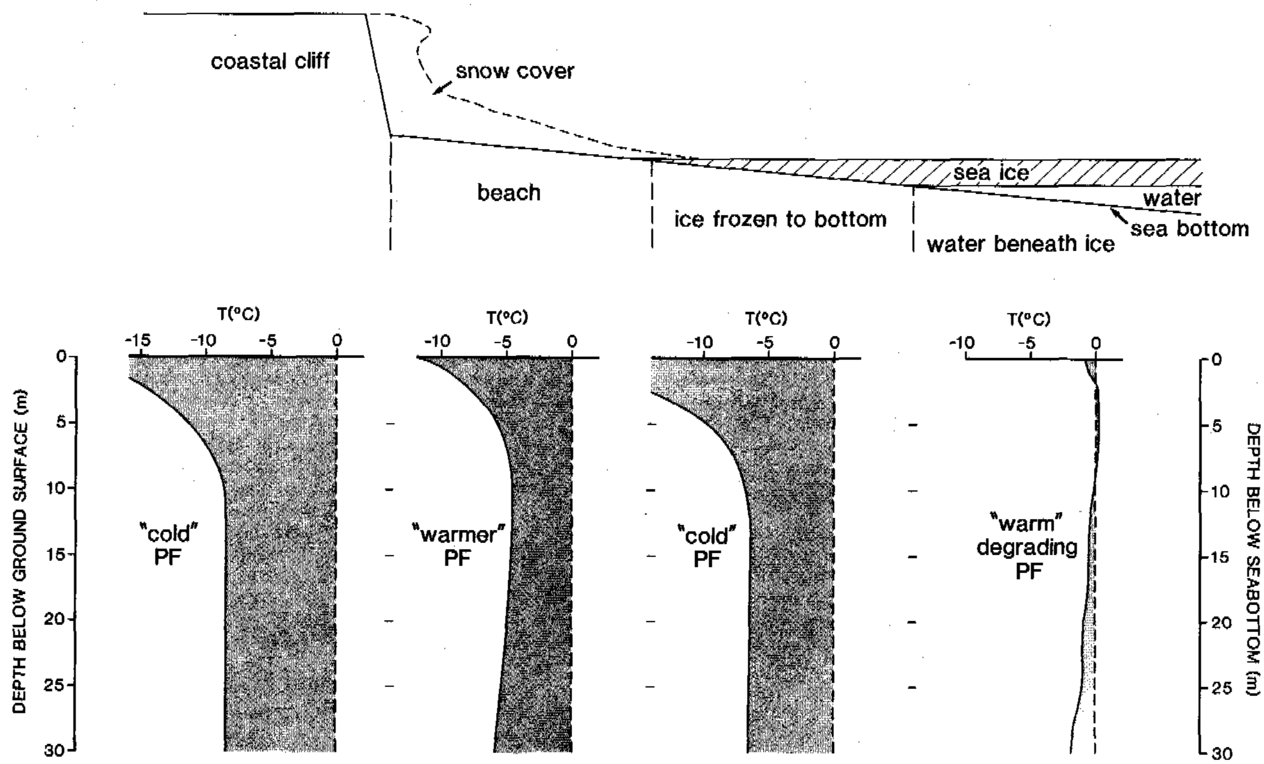

Figure 5 Diagrammatic model of the shoreline zone, along with wintertime sub-bottom temperature profiles from four holes drilled along an onshore-offshore line of the Beaufort Sea coast in the Mackenzie Delta area.

Over the winter, 1.5-2 m of sea-ice forms; thus, in the shallow water areas sea-ice is in contact with the bottom, and often has only thin snow cover. The winter cold wave is readily conducted into the seabottom, producing relatively cold sub-bottom permafrost temperatures (Hunter et al., 1988). An abrupt change in the sub-bottom thermal regime occurs at the point offshore where seawater exists beneath the ice (Fig. 5).

Alaskan Beaufort Sea

An excellent summary of shoreline permafrost conditions in Alaska is given by Osterkamp and Harrison (1985). Data collected at sites from Nome to Barrow and along the Beaufort coast show extreme variability of conditions. Subsea permafrost does not occur in the Nome area, except possibly in the immediate shoreline zone in areas of rapid coastal retreat. In the Chukchi and Beaufort Seas, ample evidence exists for coastal recession and the rapid variation of permafrost temperatures from shoreline seaward. Figure 6 shows a suite of temperature logs (after Osterkamp and Harrison, 1985) for five holes drilled from the shoreline to 7.8 km offshore near Lonely, Alaska. The changing slope of the negative temperature gradients with increasing distance offshore reflects the time since submergence. Coastal retreat rates are in the order of 10 to 15 m/year.

Unlike the Mackenzie Delta area of the Beaufort Sea, where the annual cycle of warm river water dominates the thermal regime of the shoreline areas, much of the Alaskan Beaufort coast experiences colder seabottom temperatures (<0°C) year-round. Degradation of ice-bearing permafrost under these conditions is a result of the combined effects of permafrost warming and transport of salt. The rate of salt transport into the seabed is variable, and the mechanism is not yet entirely understood (T.E. Osterkamp, pers. comm.).

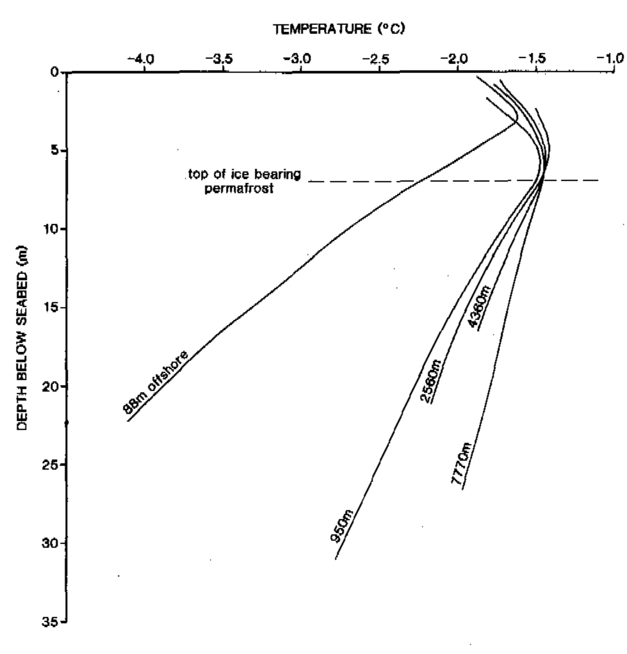

Figure 6 Temperature-depth profiles from five holes drilled from the shoreline to 7.8 km offshore in the Alaskan Beaufort Sea (from Osterkamp and Harrison, 1985).

At the Lonely site, ice-bearing permafrost occurs at depths of 6-8 m below seabottom in all holes (Figure 6). Farther offshore however, in Harrison Bay, Toimil (1985) reported ice-bearing permafrost at 2 m below seabottom in 18 m of water. This suggests that salt transport is not a major factor in the degradation of permafrost at that site. On the other hand, Osterkamp and Harrison (1985), in summarizing their data from Prudhoe Bay area, show that in the nearshore zone the transition between "unfrozen" and ice-bearing permafrost occurs in the seabottom at a temperature of -2.4°C. Salt is transported from the seabottom down to this boundary, where brine concentrations exceed that of normal seawater. Based on a coastal retreat rate of 1 m/year at this site, Osterkamp and Harrison (1985) have estimated a thaw rate of 0.3 m/year at the phase boundary. Farther offshore, ice-bonded permafrost occurs at depths between 5 and 15 m below seabottom at temperatures of -1.8°C. Seasonal freezing of the seabottom has also been noted.

Degradation of Massive Ice Bodies

Although many areas of the Beaufort coast contain large volumes of ice-rich soils with an abundance of ice-wedges and thick massive ice bodies, there are few reports of high ice-content materials in nearshore seabottom areas of rapid coastal retreat (Dallimore et al., 1988; Osterkamp and Harrison, 1985). It is possible that the majority of segregated ice bodies do not occur on land at depths substantially below present sea level, and are degraded by thermal-mechanical abrasion at the shoreline. In the Mackenzie Delta area, where there is considerable depth of thaw near the shore, any shallow sub-bottom massive ice bodies would thaw quickly. In the sub-bottom sediments of the southern Beaufort Sea, less than 1% of all permafrost samples from 86 geotechnical boreholes contained visible ice that exceeded 50% of the sample volume (O'Connor, 1984).

Mackay (1986a) studied shoreline retreat at a site near Tuktoyaktuk where massive ice is exposed in the shore cliffs. He suggests that the rate of retreat is a function of the amount of excess ice in the shoreline cliffs as well as the exposure to wave action from storm surges. Where a massive ice body did extend below sea level, an offshore bathymetric low now exists, suggesting thaw consolidation. Similar observations of deep water close to shoreline are given by Dallimore et al. (1988), for a rapidly retreating shoreline containing ice lenses and massive ice bodies.

EMERGENT COASTS AND PERMAFROST AGGRADATION

In contrast to the Beaufort Sea, the Arctic Archipelago has experienced uplift since the retreat of ice after the last glaciation (Prest, 1984). The approximate dividing line between submergence and emergence is shown in Figure 1 (J.G. Fyles, pers. comm.). The emerging seafloor areas, which may have experienced marginal permafrost conditions in the past (from cold seawater), are now exposed to lower mean annual temperatures which promote the growth of permafrost.

Taylor et al. (1983) have analysed temperature profiles from several Arctic Island well sites, and have shown that measured permafrost thicknesses and temperature gradients within permafrost can be explained by the shoreline regression estimated from emergence curves published for the region. Additional geophysical evidence for permafrost variations across the shoreline zones in the central Arctic Islands is given by Merritt (1979) and Acheson (1979). Both authors give seismic velocity and structural evidence that permafrost becomes colder and thicker inland from coastlines, and is thin, or absent, at the shoreline. Variations in seismic velocity due to differing thermal regimes can produce distortion on the seismic sections that are used to determine the stratigraphy of sedimentary rocks in petroleum exploration; and interpretation problems are often associated with the shoreline areas.

Studies of the near-surface permafrost configuration at the shoreline at Cambridge Bay, Victoria Island, have been conducted by Dyke (1987). Four sites with differing shore slopes were examined. Figure 7 shows the temperature-depth profiles in mid-summer at the high-tide limit of the four sites. The results clearly illustrate the influence of the ocean as a laterally encroaching heat source.

Dyke (1988) has also examined the growth of permafrost in the intertidal zone of Hudson Bay at Churchill, Manitoba, where the isostatic uplift rate is estimated to be 1 m/100 years. This study indicates that permafrost is aggrading in the upper portion of the tidal zone, and is modified at the shoreline by the insulating effect of vegetation. A predictive model involving seasonal freezing and thawing indices has been successfully applied to the observations.

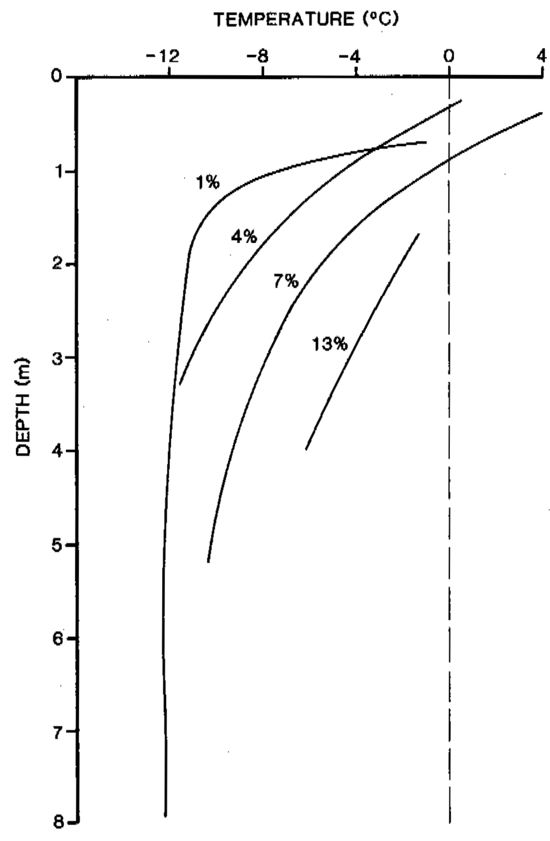

Figure 7 Temperature-depth profiles obtained in late July at four high-tide locations at Cambridge Bay, Victoria Island. The shore slopes at these sites varied from 1 to 13% (from Dyke, 1987).

SUMMARY

The coastlines of northern Alaska are experiencing coastal recession and degradation of pre-existing permafrost on the continental shelves. Salt transport into the sediments at the shoreline results in a freezing point depression; however, ice-bearing permafrost appears to occur at shallow depths farther offshore in some areas. The Canadian Beaufort Sea coast in the Mackenzie Delta area is also experiencing coastal recession; but with permafrost degradation at the shoreline strongly influenced by the warm freshwater nearshore zone resulting from the Mackenzie River outflow. No data are available for the northeastern Beaufort sea coasts of Banks or Prince Patrick Islands. However, in these areas subsea permafrost is expected to be degrading slowly without the thermal influence of river outflow.

The eastern Arctic, the Arctic Archipelago and Hudson Bay areas are experiencing isostatic uplift, and permafrost is presently aggrading at emerging shorelines. Thermal effects can be observed for several kilometres inland from such coasts. In contrast to the western Arctic mainland, observational data are sparce for the Arctic Islands and eastern Arctic mainland, and future research should be directed towards understanding the nature of permafrost at their emerging shorelines.

ACKNOWLEDGEMENTS

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PERMAFROST AND COASTAL PROCESSES

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SYNOPSIS Coastal erosion in the Arctic has recently gained increased attention from both scientists and engineers. Much of this attention has focused on those processes that are unique to environments dominated by permafrost. Within the zone of permafrost, mechanical energy is joined with thermal energy in producing thermal erosion along riverbanks and coastal cliffs. The relative importance of these two energy sources is related to a number of factors including the quantity of excess ice within the unconsolidated deposits being eroded and the intensity of ocean waves.

INTRODUCTION

The coast and its shoreline have been important to many of those people who have occupied and worked in the Arctic. Whether they be seal- or whale-hunting Eskimos or oil field developers, the shoreline serves as both bridge and barrier. The recent increase in the rate of development of the Arctic has brought with it an increase in the emphasis placed on the acquisition of information about the coast and its dynamic character.

The objective of this paper is to provide a summary of the nature of the coastline and of the processes involved in its modification with emphasis being placed on permafrost as a conditioning factor. The discussion is centered mainly on the coast that stretches from the Bering Sea to the Canadian Archipelago, a type of coast that is quite similar to a large proportion of the coast present on the opposite side of the Arctic Ocean in Siberia.

Details on the thermal characteristics of permafrost, on glaciation, on rebound, and on a number of other geophysical parameters are presented in detail in the companion paper by J. A. Hunter. These topics are pursued in this paper only where deemed necessary for clarification of the specific topic being discussed.

THE NATURAL SETTING

The coastline of the Arctic is long and varied. Using the 1:5,000,000 Map of the Arctic Regions (American Geographical Society) as a base, it has been determined that (excluding the Bering Sea and Hudson Bay) it is about 82,000 km in length (Walker, 1982).

Geology, morphology, and hydrology

The coastline circumscribing the Arctic Ocean is dominated by three physiographic types: Precambrian shields, fold mountains, and flat-bedded plains and plateaus. The Arctic Lowlands portion of this coast was the topic of a recent paper that appeared in Geomorphic Systems of North America (Graf, 1987). One of its major divisions, the Arctic Coastal Plain, contains the coastline the continental portion of which receives major emphasis in this report (Fig. 1). This coastal plain extends from the locations where the Brooks Range forms the coast in northwest Alaska to Meighen Island near Ellesmere Island in the Canadian Archipelago. It ranges from 15 to 150 km in width and has two major (the Mackenzie and Colville) and many smaller rivers that cross it. "West of the Colville River, the Arctic Coastal Plain is a marine-abraded surface mantled by marine, fluvial, eolian, and lacustrine deposits,... East of the Colville River to near the Mackenzie Delta, the coastal plain is characterized by coalescing alluvial plains and fans, with pediments increasingly common eastward along the mountain front... On the mainland east of the Mackenzie River, the coastal plain...has been interpreted as consisting primarily of a Pleistocene fluvial and deltaic plain... On Banks Island, the Arctic Coastal Plain is characterized by low rolling hills...from Prince Patrick Island to Ellef Ringnes Island [it] is low, flat, and uniform..." (Carter et al., 1987).

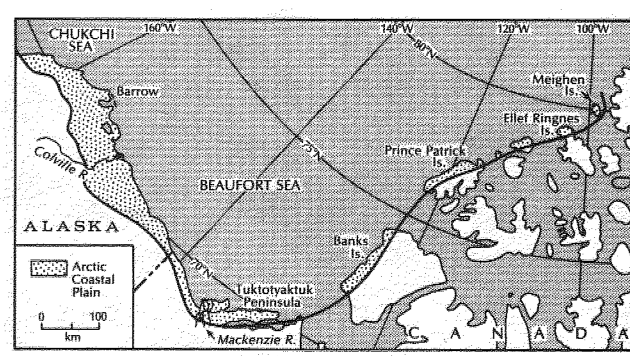

Fig. 1. The Arctic Coastal Plain (From Carter et al., 1987)

The coast (topically and regionally) emphasized in this report can be considered to be of the "soft" type, i.e., it is composed of unconsolidated mineral and organic material. Its "rock-like" nature stems from the consolidation that results from bonding by the ice that forms within the sediments.

Climate, the ocean, and sea ice

Much of the uniqueness of coastal forms and processes in the Arctic is due to the region's climate especially temperature. Low temperatures are responsible for the existence of permafrost, for the formation of lake, river, and sea ice, and for the maintenance of a long-lasting, albeit generally thin, snow cover.

Temperature is thus one of the main factors determining the seasonality of coastal processes. As long as the shoreline is protected by shorefast ice, wave erosion and nearshore longshore transport is eliminated; as long as snow drifts protect coastal bluffs (Fig. 2), thermal erosion is negligible; and as long as the active layer and rivers and streams are frozen, deltaic sedimentation is minimal. The major exception is the Mackenzie River which, like the large Siberian rivers, carries water beneath its ice cover throughout the winter (Mackay, 1963). Although the length of the period and the specific time of its initiation and cessation vary, it lasts most of the year.

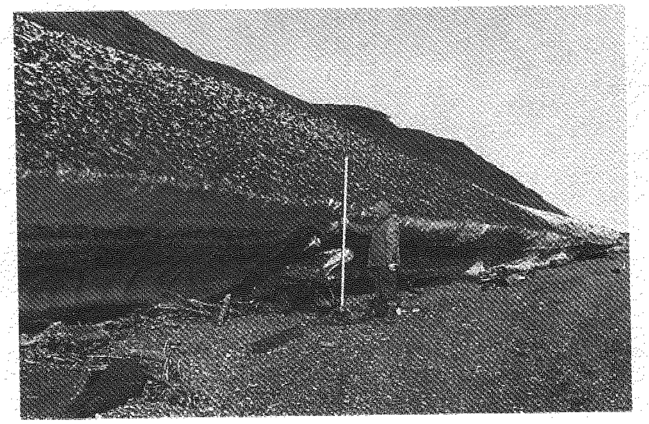

Fig. 2. Remnant snow bank in late June, Shingle Point, Canada showing results of wave action (H J Walker photo)

Winds along the coast tend to be persistent and strong. Although throughout the year prevailing winds are from the northeast, over much of the Beaufort Sea strong winds from the southwest and west frequently occur, especially during summer and fall. Under such wind systems storm surges develop and aggravated coastal erosion follows as happened in September 1986 along the Chukchi Sea coast of Alaska (Fig. 3).

Although extreme storm surges have a recurrence period of 25 to 50 years (Reimnitz and Maurer, 1978) lesser ones happen in the Chukchi and Beaufort seas regularly during the late summer and fall. Severe surges occurred in 1963 (Hume and Schalk, 1967), 1970 (Reimnitz and Maurer, 1978), and 1986. The heights of such storm surges have frequently been determined by the position of driftwood. Reimnitz and Maurer (1979) reported that the 1970 storm deposited a line of driftwood 3.4 m above sea level 5000 m from the shore. Although other reports of such high levels have been reported, Mackay notes that such extreme heights are likely the result of lodging by "...wave action, swash effect, and run-up on slopes rather than by a mean sea level rise..." (1986).

The fluctuation in the position of the ice front during summer and fall is dependent to a large extent on the direction, intensity, and duration of the wind. The position of the ice front is critical to coasts for two reasons. If winds are favorable for maintaining a wide shore lead, waves may become sizable because of the long fetch provided by the open sea. In contrast, winds blowing toward shore move pack ice in that direction decreasing fetch and dampening waves (McCann, 1973). Such ice movement, when continued to the extreme, results in ice pile-up and ice override or ivu (the Eskimo term for ice override). Both types of event can modify the beach and the bluffs behind the beach if sufficiently strong (Fig. 4). Ice movement inland of 185 m from the high-water line has been recorded on Somerset Island, Canada (Taylor, 1978). Harper and Owens found evidence of 11 override events along the Beaufort Sea coast and nine along the Chukchi Sea coast from an examination of extant air photos (1981).

Although sea ice moving on shore rearranges shore materials, creating beach ridges, little permanent change in beach form or position occurs. Shore ridges are destroyed rapidly by

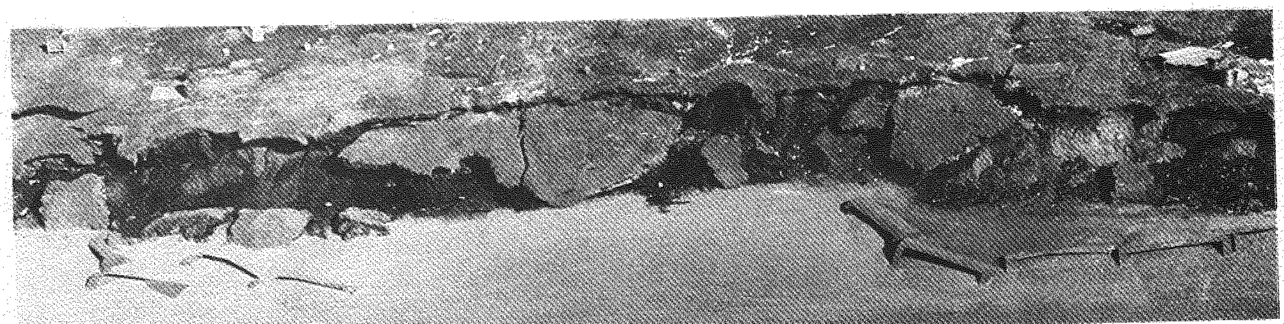

Fig. 3. Coastal erosion, collapsed blocks, and disarrangement of tar drum seawall, Wainwright, Alaska (H J Walker photo)

swash once the ice melts (Hume and Schalk, 1967).

From the standpoint of coastal development, waves, longshore currents, and tides are all important. Along the Arctic coast of Alaska and Canada astronomical tidal ranges are low, longshore currents tend to be slow, and low-energy waves predominate. However, under certain wind conditions as mentioned above, waves may be substantial especially when generated at times of high tide and low atmospheric pressure. Under such conditions wave heights have reached six meters at Barrow (Wiseman et al., 1974). Longshore currents are also accentuated during storm conditions and may result in major amounts of material transport. Hume and Schalk (1967) wrote that the storm of October 1963 "...moved more sediment in a few hours that would normally be transported in 20 years."

Fig. 4. Sea ice impact at Barrow, Alaska, 1985 as result of small ivu (H J Walker photo)

PERMAFROST, THE ACTIVE LAYER, AND GROUND ICE

Permafrost, which by definition is dependent on temperature only, may or may not have moisture present along with the mineral and organic matter of which the ground is composed (Muller, 1947). The amount of water present in the system is dependent on regional and local topography, drainage patterns, and the texture and hydraulic conductivity of the encompassing materials as well as precipitation.

Although the sine qua non of permafrost may be temperature, its actual distribution is also affected by geology, hydrology, topography, and biota (Polar Research Board, 1983). When conditions are favorable, permafrost is continuous except under those water bodies (lakes, rivers, the sea) that do not freeze to the bottom during winter. Even in such cases, relic permafrost may be present at some depth beneath the bottom of the water body as, for example, in the case of subsea permafrost.

The entire land surface of the Arctic Coastal Plain, including the immediate coast is underlain by permafrost. Its presence and the presence of the many forms that are associated with it have a major impact on the geomorphic processes that operate along the coastline.

Included in the suite of permafrost related forms of relevance to coastal morphology and processes are various types of ground ice including pore ice, segregation ice, and ice wedges), ice-wedge polygons, pingos, the active layer, and thermokarst features (Heginbottom, 1987).

The form and quantity of ice in the ground that forms the coast helps determine what happens upon coastal retreat. The amount of ice varies greatly from nearly zero as in the case of the "dry permafrost" mentioned above to nearly 100% in very localized situations such as wide ice wedges, ice lenses, and massive ice sheets. Much of the tundra bluff coastal area has values of 40 to 60% by volume in the upper few meters the bulk of which tends to be pore ice (Brown, 1967, Heginbottom, 1983). Ice wedges, in the areas where well developed, average 10 to 20% of the total volume of the upper five meters of permafrost (Harry et al., 1985). The vertical distribution of ice also varies; most of it is present near the top of the permafrost (Figs. 5 and 6).

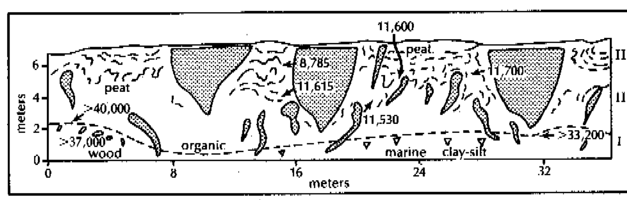

Fig. 5. Sketch of bluff at McLeod Point showing three separate systems of ice wedges (From Black, 1983)

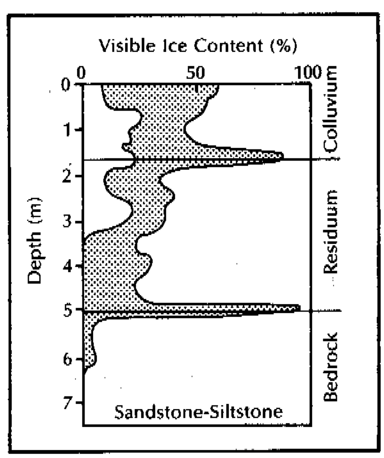

Fig. 6. Example of the distribution of ice content with depth at Melville Island, Canada (From Heginbottom, 1983)

The spacing of the ice wedges, which determines the size of the ice-wedge polygons, is important in bluff erosion as is also the angle at which they intersect the bluff. The size of wedges and the angle of intersection control the amount of ice at the bluff edge at any one time. Near King Point, Yukon Territory, Harry et al. (1985) examined a total of 50 ice wedges (80% modern) which are part of a contiguous net. The modern wedges had an "apparent" spacing ranging from 15 m to near zero where two ice wedges intersected at the bluff face.

The width of the wedges ranged up to 10 m in the case of those that were exposed at the bluff in oblique section. Such variability in spacing and width is typical of exposures that cut across ice-wedges in ice-wedge polygons. Mackay calculated that the true width of ice wedges in ice-wedge nets that have equally sized ice wedges is 0.64 times the apparent width (Mackay, 1977).

Among the most distinctive features of the coastal plain are the numerous thaw lakes that dot its surface. These lakes, many of them oriented, are frequently tapped by coastal erosion. Wiseman et al. (1973) note that "Lagoons backing the larger barrier chains appear to have originated through the erosion and coalescing of thaw lakes" and Aré (1980) writes that "...on the shores of the Bykovsky Peninsula there are isolated lagoons which according to morphological signs are flooded thermokarst basins."

PERMAFROST AND COASTAL PROCESSES

Bluff retreat along arctic coasts, rivers, and lakes has received much attention in recent years and for several reasons. It is often conspicuous, often intensive, has distinctive processes, produces unique forms, and impinges on human settlements and activities.

There are several factors involved in bluff erosion in the Arctic in addition to those normally associated with such coastal retreat outside the zone of permafrost. These additional factors center especially around the amount and type of ice included in the impacted zone. Because the ice within permafrost is temperature dependent, any condition that initiates a phase change will affect erodability. From the standpoint of bluff erosion the thermal energy transferred to the permafrost by water is especially significant although that received directly from and through the atmosphere also plays a role. Destruction by thermal energy from the air and solar radiation is referred to as "thermal denudation" by Aré (1980). These two conditions, i.e. from water and air, are sufficient to cause bank retreat by themselves under the right condition. The retreating bluff must have sufficient excess ice so that when it is converted to water the remaining sediment settles to a depth below water level (Fig. 7). The extreme case, of course, is a bluff of pure ice which when melted leaves no sediment to accumulate.

If the bluff is too high or has too little excess ice for such reduction it will develop into an equilibrium slope consistent with the texture of the thawed material. Such stabilization will develop when the thawed sediment becomes thick enough within the water body to insulate the permafrost beneath it and on the subaerial portion when it is equivalent to the thickness of the active layer.

Such thermal denudation in association with water bodies seldom occurs in isolation. Even in small water bodies some water agitation results in mechanical action and adds another dimension to shoreline erosion. These two

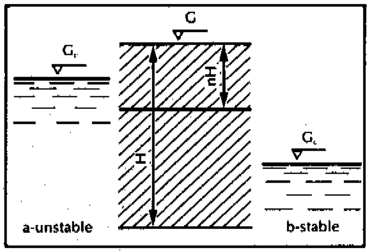

Fig. 7. The relationship between water and land in stable and unstable shores (From Aré, 1980)

processes, working together, impact river banks, lake shores, and seacoasts. Aré (1980) has labeled those coasts so affected "thermoabrasional". The mechanical element of thermoabrasion is highly varied. Along riverbanks it is mainly unidirectional although the larger the river channel the more important wind-generated waves become. Along seacoasts storm surges and storm waves become especially significant.

Both thermal denudation and thermal abrasion are summer phenomena. However, whereas thermal denudation is nearly continuous, thermal abrasion is generally irregular in timing. Along rivers it will be intense during periods of flooding and along coasts during periods of storm surges when waves extend across a beach to the cliff base.

In most coastal situations it is necessary to remove eroded material from the zone of accumulation in order to maintain bluff retreat. In this regard, what happens on the underwater shore slope is important. Thermal settling within this zone leads to deepening which in turn allows increased wave attack on shore. The rate of deepening of the subaqueous portion of a retreating coast is reduced with the accumulation of sediment which acts to insulate the bottom and prevent further thawing. In contrast, removal of this material by waves and currents enhances thawing and deepening.

Thus, coastal retreat in permafrost environments is the result of varying proportions of mechanical and thermal energy. In those cases where the permafrost is ice free (dry permafrost) mechanical erosion dominates and in those portions of the coast where excess ice is present thermal energy becomes important.

The very rapid retreat rates (see section below) that occur along many permafrost coasts have prompted many workers to place great emphasis on the thermal component (e.g., Tomirdiaro, as reported in Aré, 1980 and Reimnitz and Kempema, 1987). Reimnitz and Kempema (1987), however, have demonstrated "...that mechanical energy is more important than thermal energy for the maintenance of the dynamic equilibrium profile,..." along part of the Alaskan coast. They found that, between 1951 and 1985, thaw settlement could account for only a small proportion of the total material eroded during a coastal retreat of over 250 m (Fig. 8).

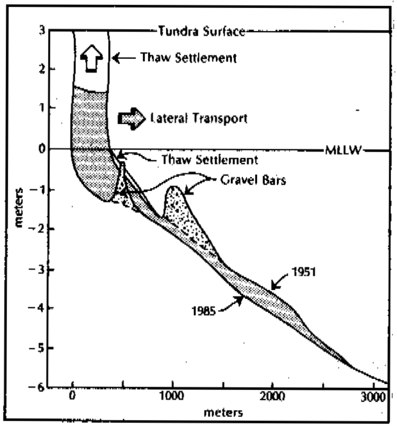

Fig. 8. Shoreface evolution, 1951-1985, northern Akaska, In submerged areas thaw settlement was nil (From Reimnitz and Kempema, 1987)

PERMAFROST, BLUFF RECESSION, AND COASTAL FORMS

Harper, in his dissertation (1978), considered four ways in which ice-rich bluffs recede; namely: surface wash, debris slides, ground-ice slumps, and thermoerosional falls (Fig. 9). "Surface wash is downslope sediment transport caused by snowmelt and summer rain runoff. Debris sliding occurs where the surficial thawed layer fails as a result of oversteepening at the cliff base" (Carter, 1987). Ground-ice slumps (Fig. 10), also referred to as thaw slumps, thermocirques, and bimodal flows, "...consist of a steep headwall, which

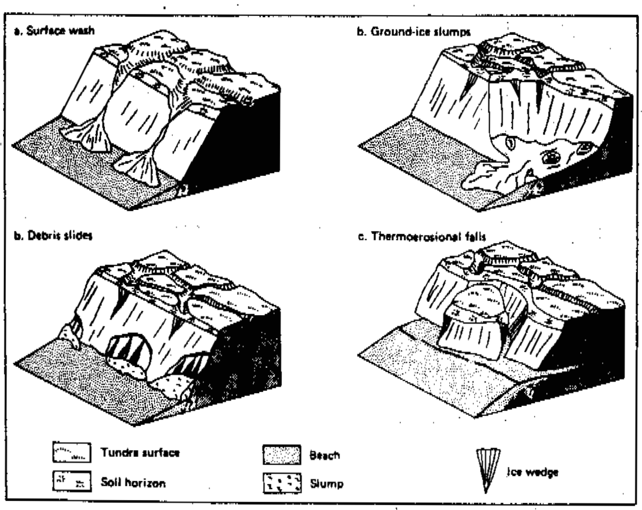

Fig. 9. Degradational processes affecting ice-rich coastal bluffs (After Harper, 1978)

retreats due to melting of ice, and a debris flow..., which slides down the face of the headwall..." (Heginbottom, 1987). Such slumping is sporadic and initiated by wave removal of debris from the base. Thermoerosional falls (Harper, 1978) or block collapse (Walker and Arnborg, 1963) provide the most spectacular form of recession in permafrost areas (Fig. 3). Such retreat is initiated by the development of thermoerosional niches along the water line (especially at those water levels that occur during river flooding and storm surges). When such niches penetrate sufficiently deep, collapse of the cornice that overhangs the niche is likely. In areas where ice wedges are present, collapse usually occurs along an ice wedge. Thus, blocks often represent part of an ice-wedge polygon (Fig. 3). Of major significance in bluff retreat is the protection such blocks provide the bank behind. Along the Colville River collapsed blocks have served this purpose for up to four and five years (Walker et al., 1987).

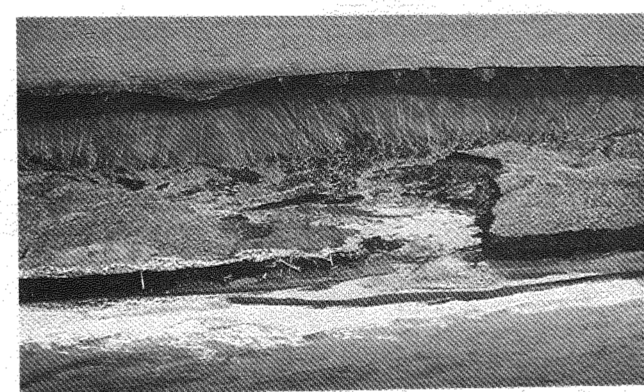

Fig. 10. Ground-ice slump, Yukon Territory, Canada (H J Walker photo)

Aré, in his definitive work on thermal abrasion (1980), provides a four-fold morphologic division of bluffs that develop along thermo-abrasional shores: sloping, vertical, truncated-sloping, and terraced (Fig. 11). As described by Aré (1980) sloping cliffs may be active or inactive and often resemble those that develop outside the Arctic. Their steepness is controlled by the angle of repose established at the time of thermal denudation and reflect the quantity of excess ice within the permafrost. When ice wedges are present, the surface may be highly irregular (Fig. 12). If an active cliff is again subjected to basal thermoerosion, removal of material at the base will gradually over steepen the lower slope and lead to the development of the truncated-sloping type (Fig. 11b). Vertical bluffs develop with thermoerosional falls. Initially, these vertical slopes will be nearly pure ice if the break occurs along ice wedges. Such ice melts rapidly and the cliff is subjected to thermal denudation. Terraced slopes on thermoterraces (Fig. 11d) develop along high coastlines in response to sporadic basal erosion working in combination with thermal denudation.

COASTAL RECESSION RATES

Many of the ice-rich bluffs that comprise much of the coast of the Arctic Coastal Plain are retreating. Coastal erosion of such bluffs varies not only with location along the coast but also greatly with time.

As Aré noted in discussing erosion rates along similar coasts in Siberia, average long-term

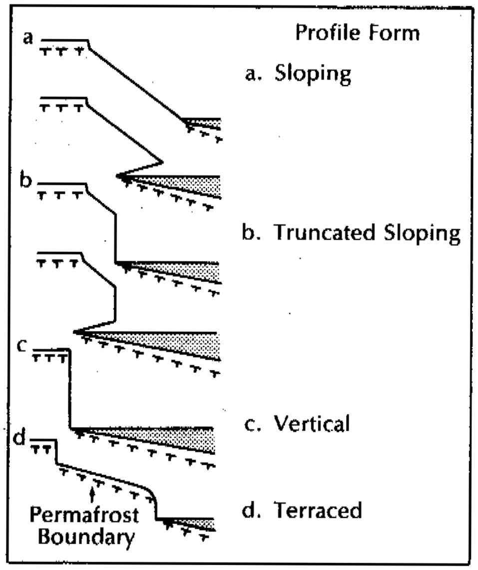

Fig. 11. The main types of profile developed on thermoerosional bluffs (After Aré, 1980)

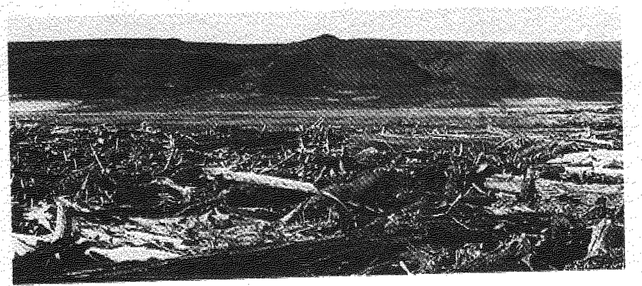

Fig. 12. Inactive coastal bluff, Blow River, Canada (H J Walker photo)

rates are quite different from those obtained during short periods of time. He reported that "...in the last 150 years, the maximum rate of thermal abrasion was observed in 1944-1946 on Semenovsky Island in the Laptev Sea to be at least 55 m per year" (Aré, 1983). Although nothing this extreme has been reported for the Arctic Coastal Plain of North America, Short et al. (1974) wrote that "In 1972, the eastern end of Pingok Island lost 40 m to tundra erosion,..." With major storms, such as that in the Chukchi Sea in September 1986, massive erosion can occur.

Average rates of retreat have been calculated by a number of researchers for various parts of the Arctic Coastal Plain. Most of the calculations tend to be for relatively short periods of time and many researchers have used different methods. Because ice-wedge polygons are relatively small features that are distinctive, photographs of them can be used to obtain accurate erosion rates. Unfortunately, the earliest vertical air photographs stem from the late 1940s, although some locations are documented on oblique photographs from earlier times.

One of the earliest calculations was made by Leffingwell who based his measurements on the retreat of the shoreline at Brownlow Point (east of Flaxman Island) in relation to an ice cellar that had been dug in 1901. By 1907 the ice cellar was being destroyed, but only after the shore had retreated by over 50 m for an average of about nine meters per year.

Using photographs as the basis of calculation, Mackay (1986) determined that the coastal recession at a massive ground ice site just west of Tuktoyaktuk was 350 m during the 50 year period between 1935 and 1985 for an average of seven meters per year (Fig. 13).

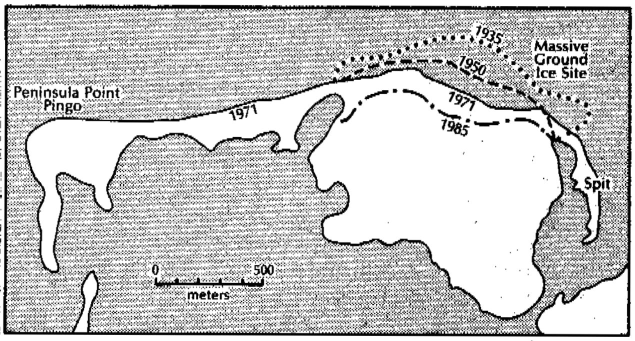

Fig. 13. Fifty years of coastal retreat at a massive ground ice site near Tuktoyaktuk, Canada. Drawn from photo in Mackay (1986)

Such rates are apparently typical of the most favorable erosional locations along the coast (Hartwell, 1972, Lewis and Forbes, 1974). Rates for most of the coast however, are more on the order of one to three meters per year (Table I and Fig. 14).

Although the above discussion has concentrated on the retreat of bluffs exposed directly to the sea, thermoerosion also occurs in lagoons, lakes, and along rivers. Erosion in Elson Lagoon (east of Barrow) was studied by Lewellen (1970). Elson Lagoon, up to 10 km in width, has low, ice rich, bluffs along most of its landward edge. Bluff retreat varies up to four meters per year in the lagoon proper. However, opposite a seven kilometer break in the barrier islands, the erosion rate is about 10 m per year. Lewellen noted that within the lagoon there was a correlation between rates of erosion and length of fetch (1970). Apparently differential erosion within lagoons results in the development of cuspate points, points that Wiseman et al. (1973) suggest reflect structural lineaments within the coastal plain east of Barrow.

Deltas, as coastal forms, present yet another example of bluff erosion. Unlike in the case of open coasts and lagoons, the water involved in thermoerosion is unidirectional in movement. That is not to say, of course, that waves generated on flowing river water are not important. In the case of large rivers such as the Mackenzie and Colville, wind generated waves can aggravate the erosional process (Walker and Morgan, 1964).

TABLE I
Coastal Retreat Rates for the Arctic Lowlands*

Location	Maxima (m/yr)	Mean (m/yr)
Chukchi Sea coast		
Near Barrow	3.9	1.9
Near Barrow	3.5	2.2
Barrow to Peard Bay	1.5	0.3
Alaskan Beaufort Sea coast		
Near Barrow	4.5	2.1
Oliktok Point	---	1.4
Near Prudhoe Bay	3	1-2
Barrow to Harrison Bay	13.4	6.3
Smith Bay to Prudhoe Bay	18	2.5
Barrow to Demarcation Point	20.5	3.0
Canadian Beaufort Sea coast		
Kay Point	---	1.3
Yukon Territory	---	1
Yukon Territory	2	1
Yukon Territory	5	---
Tent Island, Yukon Territory	27	15
Garry Island, N.T.	7.3	2.3
	1.8	1.2
Pelly Island, N.T.	13.2	6.3
Hooper Island, N.T.	2.7	1.5
Pullen Island, N.T.	---	9.2
Tuktoyaktuk Peninsula	5-8	---
Tuktoyaktuk Peninsula	6-9	---
Tuktoyaktuk Peninsula	---	4
Tuktoyaktuk Peninsula	8	---
Banks Island, (Sachs Harbour), N.T.	>4	---

*Modified from Carter (1987). See Carter, p. 603 for sources which have been deleted here because of space limitations. N.T. = Northwest Territory.

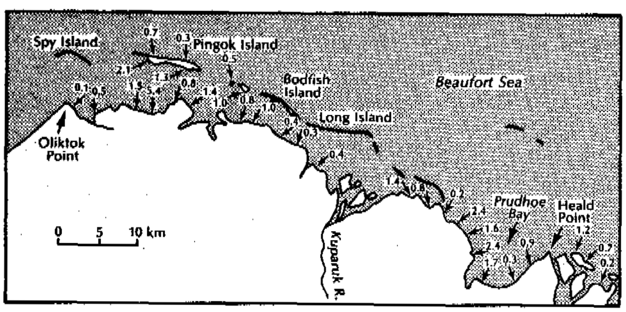

Fig. 14. Coastal retreat rates in m per year (After Rawlinson, 1983)

Rates within the Colville River Delta vary just as they do along the open coast. Cut banks and the heads of mid-channel islands are especially vulnerable (Walker, 1983a, 1983b, Walker and Arnborg, 1963). Long-term (1949-1987) rates in the tundra bluff near Nuiqsut are just over one meter per year, i.e., an amount that equals the width of the river at this location every 100 to 120 years.

CONCLUSIONS

Natrual processes along coasts composed of permafrost frequently cause rapid retreat of the shoreline. Most such retreat is the result of thermal and mechanical energy working in combination, the relative proportions of which are related closely to the amount of excess ice in the eroding system.

Recently, along some segments of the arctic coast, man has become a factor in coastal modification. By utilizing beach materials in his construction projects, he has aggravated coastal erosion as, for example, at Barrow and Wainwright, Alaska, in airport construction. To date, only small scale efforts (Fig. 3) have been attempted at reducing coastal retreat.

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MAINTENANCE OF A RAILWAY GRADE OVER PERMAFROST IN CANADA

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SYNOPSIS

The Hudson Bay Railway is approaching 60 years of operation over permafrost in northern Canada. Thaw-Settlement of the embankment and distortion of bridge decks by frost heave have been perpetual problems. This paper describes an on-going program to develop a cost effective means of stabilizing the line. Two test segments, covering 7 km of route, have been upgraded by bank widening and installation of 400 heatpipes. Ground penetrating radar was used as a survey tool to detect permafrost transitions for accurate location of the heatpipes. The sites will be monitored until 1990. The data will allow evaluation of this technique for a much broader application on the railway.

INTRODUCTION

The Canadian "north" comprises about 4.5 million sq. km or about half of our land mass. The presence of permafrost affects almost all civil engineering structures, with the greatest impact on linear transportation facilities such as railways, roads and pipelines. The demand for transportation facilities has been constantly increasing as new resources are found and Northerners are no longer content living in isolation. The high cost of construction, difficult access and harsh environment has fostered innovation among practicing engineers.

Basic design methodology for dealing with permafrost was developed in Canada in the 1950's and 60's. The 1970's saw quantification of some of the more difficult technical challenges faced by designers. Major initiatives in resource development led to improved methods of analysis of the ground thermal regime. Contributions were made to our predictive capability for creep deformations in frozen soils and to the illusive prediction of frost heave displacements. Much of the research and development conducted in Canada during the past decade was driven by a perceived immediate need to solve these more complex technical problems in order to safely produce oil or gas from the arctic.

Permafrost research and development occurred at a hectic pace during the decade that preceded the last International Conference on Permafrost. In contrast to this, the past five years have been a period of application and refinement of new technology. In some respects, the past five years can be referred to as an era of maintenance.

This paper describes some recent Canadian experience with maintenance of the Hudson Bay Railway; a pioneering facility currently operating over permafrost in Canada. On-going research to find a method for dealing with thaw-induced embankment settlement is summarized and the challenges posed by simple stream crossings are described. New technology is being applied to very old problems in the hope of reducing high maintenance costs and improving the operational safety. This quest for new and innovative solutions is vital to the very future of the facility.

THE HUDSON BAY RAILWAY

The Hudson Bay Railway was the first major transportation facility to be built over permafrost in Canada. It provides the most direct route for grain from Western Canada to a seaport which is located on Hudson Bay at Churchill, Manitoba. The line extends northeast from The Pas, Manitoba 820 km to the northern terminus of Churchill. Permafrost grades from isolated and sporadic at the south end to continuous at the north. The topography is subdued with much of the landscape either post glacial lake bed or marine lowlands that have emerged as recently as 10,000 years ago. The climatic and drainage conditions that have prevailed since deglaciation have resulted in widespread accumulation of peat.

Permafrost is present in elevated peat landforms or "peat plateaus" which have a surface covering of sphagnum mosses, lichens and black spruce. The plateaus become extensive north of the community of Gillam where permafrost underlies more than 50 percent of the landscape. Between the frozen plateaus are unfrozen water-saturated fens in which ponds and streams are common. The fen-plateau boundaries denote permafrost transitions that appear to be stable in their natural environment. However, it is hypothesized that they in fact respond to minor changes in microclimate, snow accumulation and drainage to such an extent that permafrost aggradation and degradation processes can co-exist (Thie,1974). The permafrost peatlands provided a particularly challenging environment for planning and construction of a railway to Hudson Bay at the beginning of the century.

Construction of the line began in 1917 but it was not complete until 1929. Progress was interrupted by World War I and by a re-evaluation of of the terminal site resulting in a substantial shift northward to the port of Churchill. This added about 200 km to the length and forced the route over the Hudson Bay Lowlands, one of the most extensive permafrost peatlands in the world. A description of the grade construction techniques is provided by Charles (1959). He comments on the importance of route location, avoidance of cuts, placement of fills over frozen peat and the stability of water crossing structures; both bridges and culverts. This historic paper documents 30 or more years of early Canadian experience with embankment construction over permafrost.

The line has been in continuous operation since 1929. However, with passing time, quality of the track has deteriorated, maintenance costs have increased and operational hazards are now a real concern on the system. Even with annual expenditures on maintenance up to 3 times normal costs for western Canadian railways it has barely been possible to keep pace with grade deterioration. The situation is further aggravated by improvements in rolling stock that demands a better class of track structure to carry greater axle loads at higher speeds. The operational challenge is primarily due to the presence of permafrost with the widespread occurrence of peat being a secondary factor.

THAW-SETTLEMENT

Thaw-settlement of the embankment at a discrete location has been termed a "sinkhole" by maintenance personnel. Sinkholes are widespread within discontinuous permafrost and there are indications that they are spreading north, well into the region where permafrost in normally considered continuous. Most sinkholes require 100 to 150 mm of fill each year to maintain the track in a serviceable condition. Sinkholes invariably form at the transitions between permafrost peat plateaus and the adjacent fens. A section of track across the permafrost peatlands is shown in Figure 1 and a typical sinkhole is

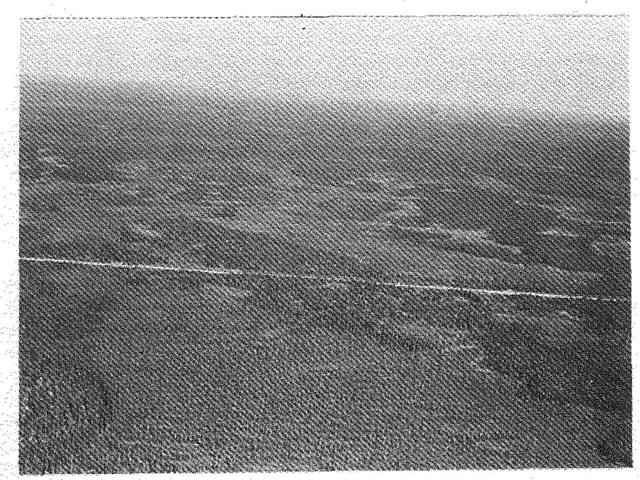

Figure 1 Railway crossing discontinuous permafrost peatland north of Gillam, Manitoba

An hypothesis that explains the occurrence and distribution of sinkholes is reproduced as Figure 3 (Hayley et al.). Progressive lateral thaw of an unstable, near vertical, permafrost boundary causes the sinkhole to slowly migrate into an otherwise stable frozen peat plateau. The key to restoring stability of the embankment resides with stopping the retrogressive thaw. Vertical heat pipes, inserted through the fill into the thawing subgrade, were shown to be effective at cooling the transition after a trial installation in 1978 (Hayley et al,1983). In order for this technique to be adopted for large scale use along the route it must be shown to be not only technically sound but economically viable. The following factors were identified as critical to adoption of heat pipes as a route stabilization technique.

a) Characterization of sinkhole distribution and frequency along the route.

b) Development of a rapid and effective means of locating the permafrost transitions in order to correctly position the heat pipes.

c) Adoption of a generic design that is relatively independent of site-specific conditions.

d) Development of an efficient, rail mounted, means of installing the heat pipes.

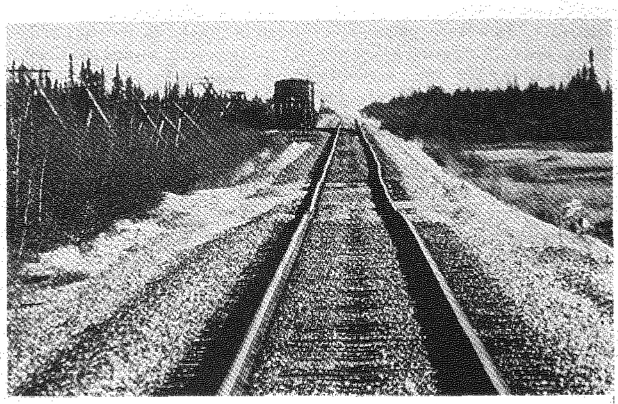

Figure 2 Typical sinkhole

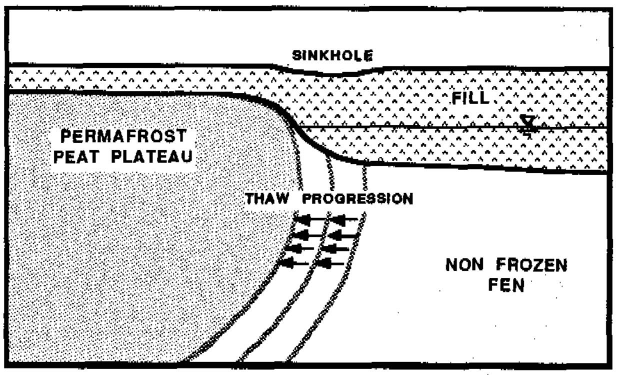

Figure 3 Sinkhole formation hypothesis

A thorough evaluation of the 600 km route over permafrost was conducted by review of track maintenance records, airphoto examination and geophysical methods. The northern 100 km and the southern 200 km are relatively free of thaw-settlement. An approximate distribution of sinkholes by frequency for the middle half of the route is shown in Figure 4. Seven hundred active sinkholes were identified and characterized in this study.

The distribution of sinkholes can be directly related to variations in organic soil cover and climate. The sinkholes can also be broadly categorized as simple or

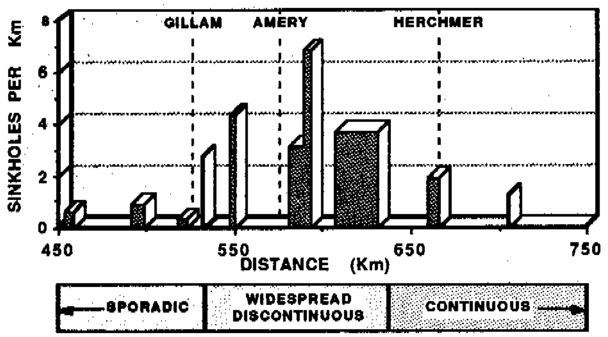

Figure 4 Frequency of sinkhole occurance relative to the distribution of permafrost.

complex. Simple sinkholes are characterized by a short, abrupt dip in the track. They typically have 2 to 3 transitions from frozen to non-frozen subgrade soils. In contrast, complex sinkholes are often as long as 300 m, and are underlain by 4 to 7 permafrost transitions. The complex sinkholes represent a more advanced stage of permafrost degradation and are common to the southern half of the route. A closely spaced group of simple sinkholes impedes the progress of train traffic. However, it is the long complex sinkholes that pose the greatest operational hazard because differential settlements are more severe both along and across the track.

Application of Ground Penetrating Radar

Effective use of heat pipes requires a reliable method for determining the location of permafrost transitions within individual sinkholes. Experimentation with ground penetrating radar on the Hudson Bay Railway has confirmed that it is both fast and effective at locating the permafrost transitions for soil conditions particular to this route. Ground penetrating radar has been used experimentally for some time to differentiate between frozen and thawed soils (Annan and Davis, 1976) or to characterize permafrost by ice content (Kovacs and Morey, 1985). However, there is no precedent for application of this technology to collection of route data for design purposes.

The first radar survey was conducted along 77 km of route during the winter, 1985. The antenna was towed behind a vehicle at speeds ranging from 5 to 7 km/hr. High frequency pulses (120 Mhz) were transmitted into into the ground and a reflection was received at the same antenna. The signal was processed on electronic equipment carried in the vehicle then printed in analogue form on a strip chart recorder. A subsurface boundary between granular and cohesive or organic soils or between frozen and unfrozen soils can normally be distinguished because there is a change in electrical properties of these materials.

Typical data from a radar trace over one of the early heat pipe test sites (1978) is shown in Figure 5. The bottom of the granular embankment can always be distinguished as can the groundwater table. Fill that has been recently placed in active sinkholes is also evident. The radar signature from the peat plateau is more intense than the return from the fen. Nevertheless, precise detection of the frozen, non-frozen interface is not particularly reliable perhaps because it is normally coincident with the base of the fill. The nature of the interface is such that a relatively steep slope is defined where there is a transition from thick fill over unfrozen peat (fen) to thin fill over a permafrost peat plateau. These characteristic features on the radar profiles are considered to be reliable indicators of the degrading permafrost boundary. This has been confirmed by comparison with geotechnical data obtained by drilling and sampling.

Reasonable success from initial field trials prompted use of radar to locate heatpipes just ahead of an installation crew and as a monitoring tool. Results of sequential survey data obtained from both summer and winter surveys have been compared to assess changes.

Prototype Heatpipe Installation

A prototype installation was completed in 1987 to test the validity of using heatpipes to stabilize a meaningful segment of track. Two, 3.5 km long route segments were chosen to represent worst case conditions on the northern and southern ends of the most troublesome portion of the route. These sites reasonably covered the range of closely spaced simple sinkholes (northern site) and long complex sinkholes (southern site). The permafrost transitions were located from ground penetrating radar traces. A generic design was developed for each of the sites based on results of the 1978 installation (Hayley et al,1983). Data from the previous installation indicated that a transition

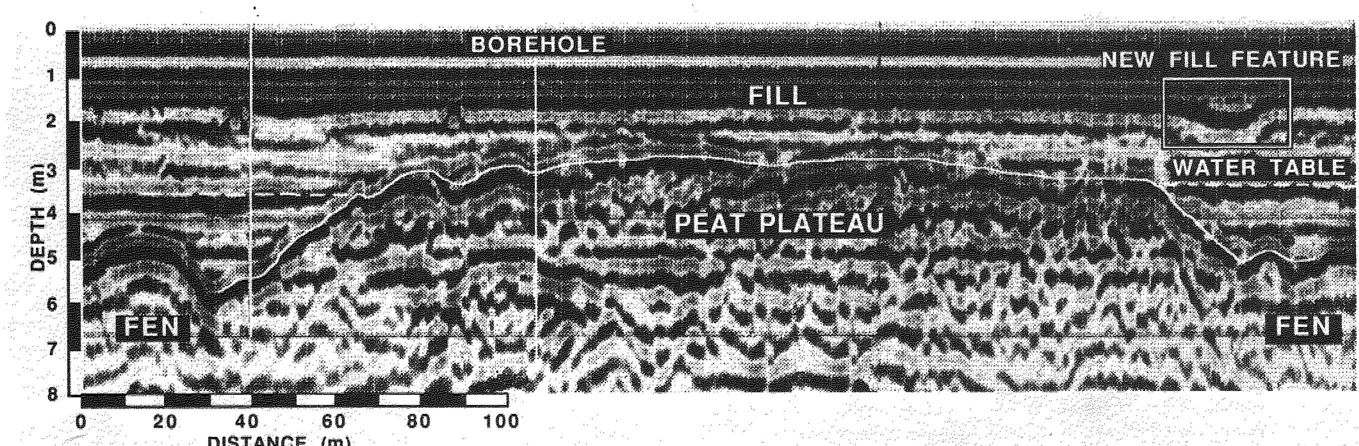

Figure 5 Ground penetrating radar trace over 1978 test site no. 4

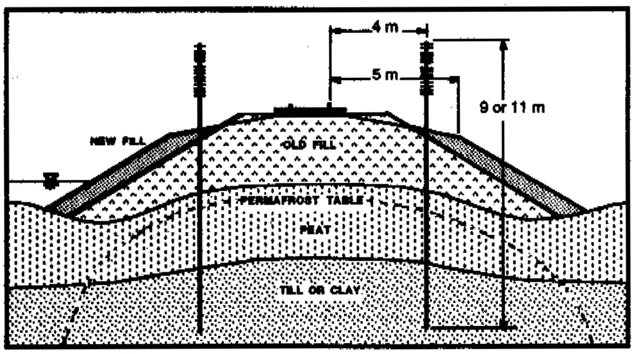

Figure 6 Heatpipe installation schematic

could be stabilized with as few as four carefully located heat pipes spaced at 4 m. with 2 on each side of the rails.

A sketch of the generic design adopted from past performance observations is shown in Figure 6. Selective bank widening was required to stabilize the sideslopes. The heatpipes were 9 m long for the northern site and 11 m long for the southern site.

This design was supported by thermal performance data obtained from two years of comparative testing at Thompson, Manitoba. The two units that are manufactured in Canada; Cryo-Anchors and Thermo-Probes, were installed at a non-permafrost research site and monitored to determine the extent of the frozen cylinder that formed around the pipes each winter. A heat transfer coefficient was back-calculated from the results for comparison with the manufacturers published data. The results compared well, resulting in acceptability of either unit for use on the railway.

Construction was initiated in September, 1987 following a competitive bidding process for both supply of the heatpipes and installation services. The heatpipes chosen were manufactured in Winnepeg, Manitoba by Arctic Foundations of Canada Inc. These units were of standard Thermo-Probe design; charged with carbon dioxide. The steel pipes were upgraded to make them more robust and to allow a free standing radiator.

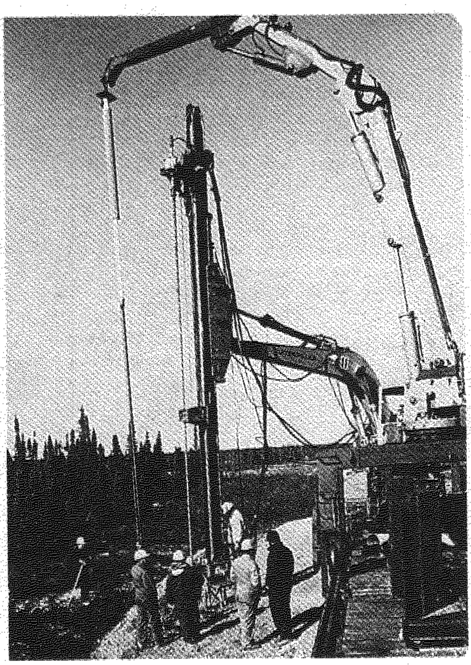

Figure 7 Auger drill and crane installing heatpipes

Figure 8 Installation crew with rail mounted equipment

A total of 400 heatpipes were installed in 26 days using a custom designed, rail mounted drilling rig. A continuous flight auger, mounted on a long mast attached to a backhoe, rapidly drilled a hole that was slightly larger than the diameter of the heatpipe evaporator (61 mm). Mud was injected through the centre of the auger to fluidize the drill cuttings and to stabilize the hole. The heatpipe was installed with a small crane immediately after the augers were withdrawn. The entire operation was conducted from two flat cars that were moved with a small shuttlewagon. The construction operation is shown in Figures 7 and 8. Installation rates varied from 1 to 6 units per hour with an average of 3.3 units per hour.

Representative segments of the test sections have been instrumented with thermistor cables and settlement survey points. These will be monitored for two years to provide a quantitative means of evaluating performance. In addition, operating and maintenance personnel will provide their own assessment of the effectiveness of the installation. A significant improvement in overall track quality throughout the 3.5 km test sites is essential before the measures can be judged a success.

STREAM CROSSINGS

General upgrading of the railway would require replacement of a number of timber bridges. The timber bridge decks are all supported by timber piles that have commonly experienced substantial frost jacking. The small bridge, shown in Figure 9, illustrates a rather severe case of bridge deck distortion caused by pile heave. The structure at km 612, shown in Figure 10, is more typical of many of the original timber bridges on the route north of Gillam. Even with extensive shimming, piles no longer bear evenly on the cross beams, as shown in Figure 11. In many cases, the piles have been cut off as they were progressively jacked out of the ground.

The crossing at km 715 provides some insight into the extent of the redesign measures that are required to provide a stable structure in this environment.

Figure 9 Bridge deck distortion due to frost heave (1978 photo)

Figure 10 Typical timber bridge at km 612

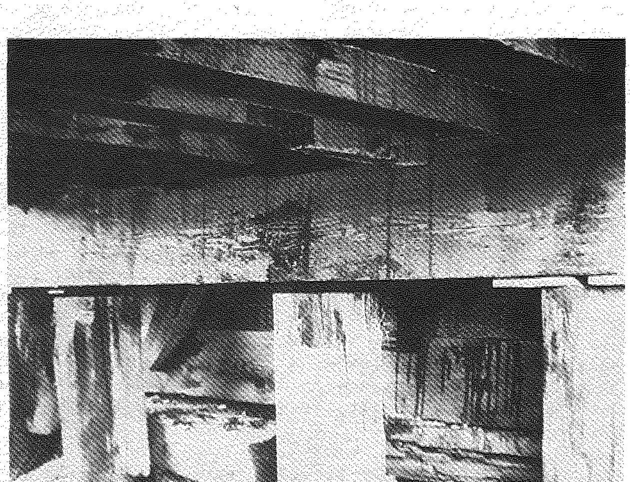

Figure 11 Gaps and shims on top of timber piles

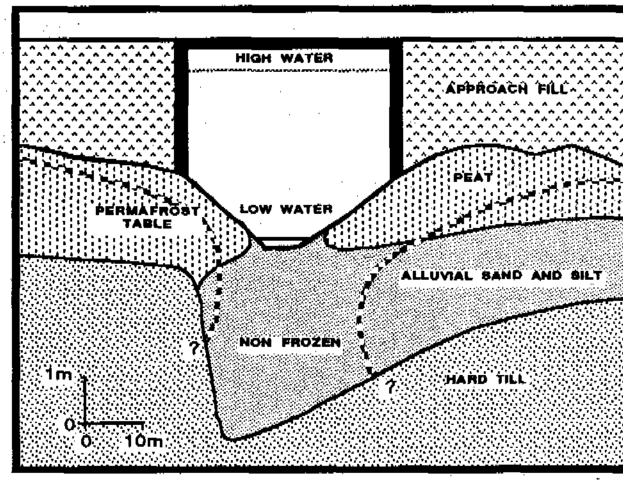

Figure 12 Site conditions at km 715 crossing

Figure 12 is a schematic illustration of the soil conditions at this location. It was constructed from four geotechnical boreholes and limited survey data. The stream is about 10 m across with a 12 m wide floodplain on each bank. The original timber structure was 34 m long, with 10 simple spans between pile supported bents. Soil conditions in the valley consist of alluvial silt and sand with a veneer of gravel overlying very hard till. Thick peat overlies the alluvial soils on the floodplain and it extends under the abutments and approach fills. The alluvial soils are only 5 to 6 m thick thus the timber piles met refusal at a shallow depth in the hard till (about 7 m).

Permafrost is not present below the active channel or on the floodplain but is present below the abutments. The water depth varies substantially throughout the year but seldom exceeds 1 m. It is usually very low in late fall (300 mm) thus the stream will freeze completely in winter, except perhaps for one small channel.

The original timber structure was replaced in 1978 with the steel bridge shown in Figure 13. There are five bents each supported by 6 steel H piles (300 by 300 mm) that were driven to depths ranging from 11 m to 12.5 m. When the bridge was inspected by the author in 1986 heave of the piles supporting the centre bent was obvious. The centre bent was 100 mm above the adjacent ones with the displacement clearly noticeable on the lower flange of the superstructure. Only the piles supporting the centre bent, located in the stream channel, had heaved (Figure 13).

The depth of seasonal frost penetration into the streambed was estimated to range from 2 to 3 m. The uplift resistance for piles driven into unfrozen alluvial soil and till is predicted to be about 565 kN, thus the average heave stress for 3 m of frost penetration probably exceeded 150 kPa. Such a high average heave stress has also been reported by Penner and Goodrich (1983) from tests on restrained steel pipe piles at Thompson, Manitoba. Their data shows the detrimental effect that a 0.5 m thick surface layer of gravel has on the adfreeze stress, increasing the average stress by a factor of about four. Such a condition usually can not be avoided when piles are driven into an active streambed.

Reconstruction of the timber bridges on the route must carefully address the pile heave potential. Clearly,

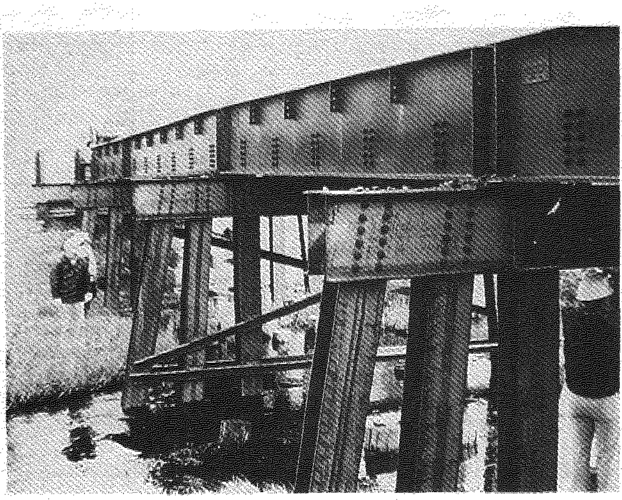

Figure 13 Steel replacement structure for Km 715 showing heave of the centre pile bent

timber piles are not a good choice for the conditions described because they cannot be driven deep enough to establish sufficient anchorage against frost heave. Steel H piles can be driven to considerable depths, however, they expose the maximum surface area per meter of length to adfreeze. Steel pipe piles provide the best compromise between drivability and heave potential. To ensure that adequate anchorage is provided against heave a penetration depth in the order of 20 m would be required. Alternative solutions to deep driven piles have been considered but rejected because they have been judged to add unreasonable complexity to the construction effort.

CONCLUSIONS

The practice of engineering for permafrost conditions in Canada has matured to the extent that we now have transportation facilities and structures that have completed their life cycle. These facilities require either extensive maintenance or reconstruction if they are to continue in operation. The Hudson Bay Railway is a good example of such a condition. Its continued viability is important to the commerce of northern Manitoba but its future is threatened by rising maintenance costs and an inability to keep pace with improvements in design of rolling stock.

Settlement of the track structure is a result of retrogressive thaw at the transitions from non-frozen to permafrost peat landforms. Research to date has shown that selective installation of heatpipes will arrest thaw-settlement at locations where conditions fit the model developed (Hayley et al.,1983). The on-going study described in this paper addresses the practical extension of the earlier work by developing a cost effective methodology for design and installation on a much larger scale. A key element in the design is a demonstrated capability to locate the heat pipes using ground penetrating radar. Success of the new installations will depend primarily on qualitative evaluation of track condition through the 3.5 km test sites by operations and maintenance personnel.

A plan to restore or upgrade the railway must also include replacement of many original timber bridges that have been distorted by frost heave. Wooden piles are unsatisfactory for new construction because they can not be driven into hard till or permafrost that underlies most sites. Experience with driven steel H piles has not been entirely satisfactory because of the severe potential for frost heave. A steel pipe pile, driven to a depth of 20 m, is required to provide reasonable assurances that stability will be maintained.

ACKNOWLEDGMENTS

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AIRFIELDS IN ARCTIC ALASKA

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SYNOPSIS Airplanes provide an important means of transportation for villages and petroleum-related activities in remote areas of northern Alaska. Accordingly this review discusses temporary and permanent airfields which have been constructed in that region, ranging from snow and ice winter airstrips to gravel and rock fills for all-season airfields, which can eventually be paved to provide permanent airports.

INTRODUCTION

Arctic Alaska can be considered remote, having only brief and limited access to the sea and very few roads. Although overland transportation has changed from tractors and sleds to vehicles with low pressure tires, and even air cushion vehicles, the principal means of transportation to many of these remote areas is the airplane. The use of the airplane in the North is not new, but aircraft have changed, to accommodate more passengers and larger and heavier freight. Construction and drilling equipment is designed and built to be transported by specific airplanes, such as the C-130. For more than a decade it has been possible to move entire oil rigs, with the associated supplies, camp and construction equipment, from one remote area to another in the Arctic, provided there are temporary airfields at both locations. Temporary airfields are therefore vital elements to air transportation in the North.

Temporary airfields in the Arctic are practical, prudent and advantageous solutions to the siting and construction constraints of a specific location. When used in petroleum exploration the location for a new airfield is primarily dictated by the proposed exploration well location. Unfortunately the well locations are selected from geophysical data, with little regard for the existing terrain conditions. Accordingly many temporary airfields differ from permanent airfields in many respects. For instance the orientation of the runway may be based on strong winter winds, if the operational period will be during the winter only. Topographic features may also be used to advantage, with due respect for cross winds that might occur only in winter storms. Temporary airfields, being private airfields (without scheduled flights), can be closed to all traffic in storms and strong cross winds. State and federal guidelines for the design of airports can also be waived in some circumstances with respect to lateral obstructions (drill rigs), and even approach obstructions. Some airfields are constructed with access only from one end of the runway, at least for large aircraft, the other end being obstructed by a hill or mountain. This is not unique to the North Slope, for there are several airfields elsewhere in Alaska that have one-way access and have been in continuous operation for more than 25 years. While every attempt is made to comply with normal airfield standards, there are situations where this would be impossible, prohibitively expensive, or require that the airfield be many miles away from the desired location. Nevertheless, all variations from standard practice are carefully evaluated by all parties, including the pilots, aircraft owners and their insurance companies. Temporary airfields are not expedient or to be otherwise construed as lacking in good engineering or workmanship, for lives and millions of dollars are involved in every takeoff and landing.

TYPES OF AIRFIELDS

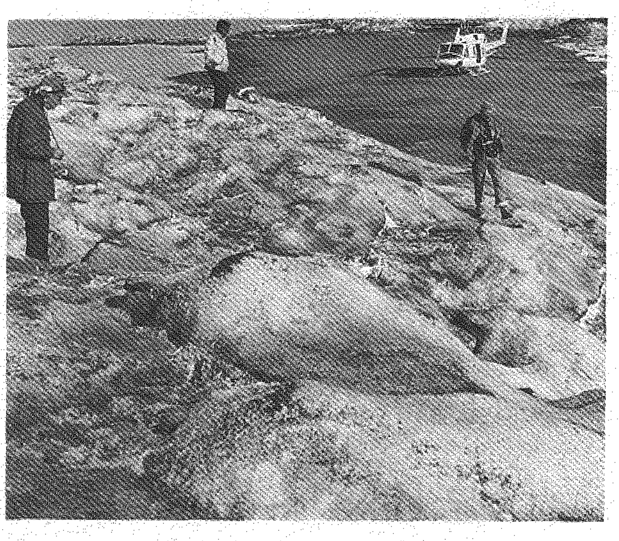
For purposes of this review the types of airfields can be grouped according to the construction material used, or their location, as follows:

A. Runways on Ice
   1. Sea ice
   2. Lakes or ponds
   3. Rivers
B. Snow or Iced Runways
   1. Compacted snow or ice
   2. Partially drained lakes
   3. Ridges, beaches and gravel bars
C. Gravel or Rock Fills
   1. Conventional gravel fills
   2. Insulated gravel fills
   3. Rock fills
   4. Dredged material
D. Combinations and Other Types
   1. Combinations
   2. Landing mats
   3. Geo grids

The above groupings do not constitute a ranking or preference; the selection of the runway type best suited for the proposed site is dictated by the site itself and the period during which the airfield will be needed. For instance a shallow exploration well, along the coast or in an area of many lakes, can be scheduled for winter drilling, wherein ice can be utilized. Conversely the planning of a deep exploration well, or other construction that might require a year or more to complete, would consider all-season airstrips using a gravel or rock fill. Many programs in remote areas also require more than one runway. A frozen lake may be used to initially move in the heavy equipment to build the gravel runway, or a short snow runway may be built close to the drill rig, while a large frozen lake, several miles away, may be used for larger aircraft. Even when drilling rigs are brought overland, on winter trails, a short temporary airstrip is usually built on ice, snow, beaches, ridges or gravel bars. Such small airstrips, for STOL type aircraft, are considered essential for obtaining small parts and supplies, crew changes and medical emergencies.

Runways on Ice

Sea ice may be utilized for temporary winter runways. These include free-floating, first-year ice along the coast, or bottom-anchored ice near the shore, in bays, lagoons or shallow waters. Site selection for such runways usually includes a search of historical information on ice conditions in the specific area to determine the extent and location of pressure ridges and previous ice movements (Bilello, 1980; Continental Shelf Data Systems, 1969). Winter storms along the open coast can destroy such airstrips, piling ice along the shore to heights of 50 ft, as shown in Figure 1. Such ice movements can also disrupt airfields on beaches, in both the summer and winter, as described by Kovacs (1983). Accordingly, where possible, such runways on sea ice are preferentially located between the coast and offshore barrier islands, in lagoons or embayments protected by sand spits or other features. Sea ice is weaker than freshwater ice and normally requires a greater thickness of ice for a given weight of aircraft, as defined in Transport Canada (1985) and U.S. Air Force manuals (1958, 1968). Sea ice can be thickened, by pumping and flooding, as is done with some freshwater ice runways, to achieve the required thickness earlier in the winter.



Many of the freshwater lakes, which abound on the Arctic Coastal Plain and even in the Foothills, have long been used for winter airstrips. These also include the small shallow lakes, which have primarily been used for the shorter runways, while the deeper and longer lakes have been successfully used for the 1600-m C-130 runways. While the water may freeze to the bottom rather quickly, at least along the shores, the floating ice thickness in the middle of the lake is often the controlling factor. The deep central region of many arctic lakes is elliptical in shape.

Lakes under consideration for a runway should be inspected from the air and the bottom profiled along the proposed alignment. Concern here is for the difference in dynamic stresses and deflections in a floating ice sheet, as opposed to the bottom-anchored ice in the shallows. Such dynamic conditions are generated when the aircraft is landing, taking off, or taxiing at high speeds (Nevel, 1979). Whenever possible one should try to park and unload aircraft on ice frozen to the bottom (Fig. 2) or limit the duration of parking on floating ice. Snow is usually removed from the runway during initial construction and subsequent operations to promote quick and deep freezing. If early use of the runway is desired, the ice may be thickened by flooding in thin lifts, using water from the deep portions of the lake. After flooding, the surface of the ice is scarified with a grader blade to provide good traction and braking. One should always check the salinity of all lakes near the coast for they could have been contaminated by previous storms.

While river ice was recognized in the early exploration programs in PET-4 as providing a convenient route for tractor trains (Reed, 1958) frozen rivers can also be used for runways. However, unlike lakes, some of the larger rivers on the North Slope experience a considerable drop in elevation after freeze-up. Further lowering of the river after freeze-up causes the ice near the shores to crack and to be left stranded. However, some reaches in the wider and deeper rivers, such as the Kuk River, can provide excellent runways (although the northern reaches of the Kuk River can be brackish). Strong under-ice currents in narrow river reaches can retard freezing, create thin spots, or prematurely thaw in the springtime, posing a threat to

Fig. 2 Aircraft on frozen lake, just north of Inigok, spring of 1978.

Fig. 3 Outline of runway in drained lake, after thawing, East Karupa No. 1 well site, 1980.

both construction and operations, and hence should be avoided.

In general, runways on ice can provide excellent airfields for short-term use, often with much less work than required for snow or iced runways on the tundra. The environmental impact of runways on ice is minimal, the biggest concern being fuel spillage. There usually is no impact on the permafrost, except at shore-based facilities, or the approach roads. Construction and operations supervisors should have experience and full knowledge of ice mechanics, hydrostatic stresses and dynamic loadings, including the thermal and load cracking of ice.

Snow and Iced Runways

Runways can be constructed directly on the tundra, in the same fashion as winter trails. Normally such runways are for STOL aircraft and are limited in length to about 760 m. Some STOL aircraft are equipped with skis, which avoids many of the problems of wheels on snow (Abele et al., 1968). The runways are usually constructed by first packing down any loose snow. Of particular interest here is to fill the voids between any and all grass tussocks. The runway can then be thickened and graded to the desired elevations by subsequent applications of snow, which is saturated with water after compaction. Harvesting sufficient snow, without causing damage to the tundra, is an important consideration and deep snow drifts along rivers, lakes or the sea coast are often the best sources. Snow fences, deployed earlier, can also be effective in providing snow for construction (Williams, 1978). However, if ice from shallow lakes or rivers is readily available, the runway should be thickened and graded with chipped ice. Chipped ice requires much less water to congeal into a frozen mass than snow, with less freezeback time (Adams, 1978a, 1978b; Fisher, 1977). Runways on the tundra can also be constructed by direct flooding, using water from nearby rivers or lakes (Mitchell, 1981).

Runways can also be constructed in drained or partially drained lakes. Such runways utilize the shallow ice frozen to the bottom in the center of the old lake bed, with compacted snow or chipped ice extending beyond this central section to achieve the desired runway length. Drained lakes usually have limited areas of aquatic grasses, as opposed to tussocks. The extent and abundance of these aquatic grasses are a function of the original depth of the lake, how long the lake has been drained, and the extent to which it has been drained. An example of such an airfield, near the East Kurupa No. 1 well site, is shown in Figure 3, several years after use. A major concern when using drained or partially drained lakes is the possibility of the emergence of a pingo under, or beside, the runway (Mackay, 1983).

Ridges, beaches and gravel bars have provided numerous runways for STOL and small cargo planes, particularly during the early petroleum exploration programs on the North Slope (Reed, 1958). While many were used only in the winter, when frozen and snow covered, some were and still are used in the summertime. Normally these runways required little initial preparation, although some gravel bars have required spot grading, filling and the removal of brush. During the summer many gravel bars are too rough and loose for regular tires and "tundra" tires must be used. However, when frozen, snow covered and iced down, they can provide a good runway. River bars and beaches normally do not involve major earthwork and are usually destroyed or rendered unserviceable by storms or floods. Brush, if present, quickly encroaches on such temporary facilities, once they are abandoned. Ridges, which are often windswept and barren, normally reflect surface damage only where the residual soils are thick and previously supported vegetation.

Gravel or Rock Fills

Virtually all permanent or temporary all-season airstrips in northern Alaska have been built using gravel fills. Airfields close to the coast have utilized beach gravels that unfortunately are uniformly graded, with well-rounded pebbles or cobbles. Accordingly the surface course must be blended with silt and sand size material to achieve a smooth, durable and high bearing capacity runway that will not rut or ravel. Gravel fills for such runways are normally placed directly on the vegetation to retain the insulating effect and latent heat retention of the near-surface organics. These fills are placed in both winter or summer. However, if done in the summer, placement is by end dumping to avoid disturbing the vegetation and permafrost, particularly along the shoulders. The initial design depth of the gravel fills normally ranges from 1.5 to 2.0 m, although some runway fills have been as shallow as 1.0 m. Such gravel surfaced runways can be graded and recompacted, during the summer months, but often lose their fine-grained soil binder over the years. Overlays of select gravels are therefore required periodically, as typified by maintenance of the Umiat airfield, shown in Figure 4. Paving of gravel-surfaced runways on permafrost is usually not considered until any abnormal thaw settlement has taken place and the runway has been regraded with additional gravel over a period of 10 or more years. To date only two airfields on the North Slope have been paved, those at Barrow and Deadhorse.

Insulated gravel airfields are employed to protect and preserve the underlying permafrost when there is limited gravel available to provide the full depth of conventional gravel fills (Lachenbruch, 1959). The design of such airfields employs 4.0 to 7.5 cm of high-density polystyrene board insulation, placed directly on frozen cut or fill sections of sand or other unclassified borrow. A minimum thickness of 30 cm of gravel is placed atop the insulation to provide the required bearing capacity, gradable surface and to protect the insulation from being crushed (Crory et al., 1978). There are three such insulated runways on the North Slope (Inigok, Tunalik and Ivotuk), the Inigok runway being shown in Figure 5. All of these insulated runways were constructed in the late winter and early spring to capitalize on the coldest ground temperatures possible.

Quarried rock can also be used for runways, even when gravel sources are within reasonable distances. The airfield at Inuvik, in the Northwest Territories, was constructed with rock fill (Johnston, 1982). An example of a temporary airfield is shown in Figure 6. This runway, at the Killik well site, had a 4.5-m-high rock fill on the north end. All rock for

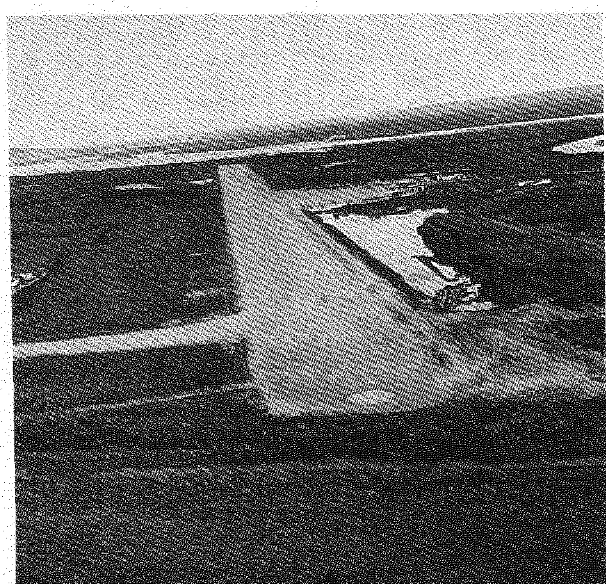


Fig. 4 Umiat airfield, originally built in World War 2, after applying new surface gravel, 1980.

Fig. 5 Insulated runway at Inigok on sandy ridge constructed with limited gravel from Colville River, 1978.

Fig. 6 Rock fill runway at Killik well site with deep fill on north end, in foreground, 1980.

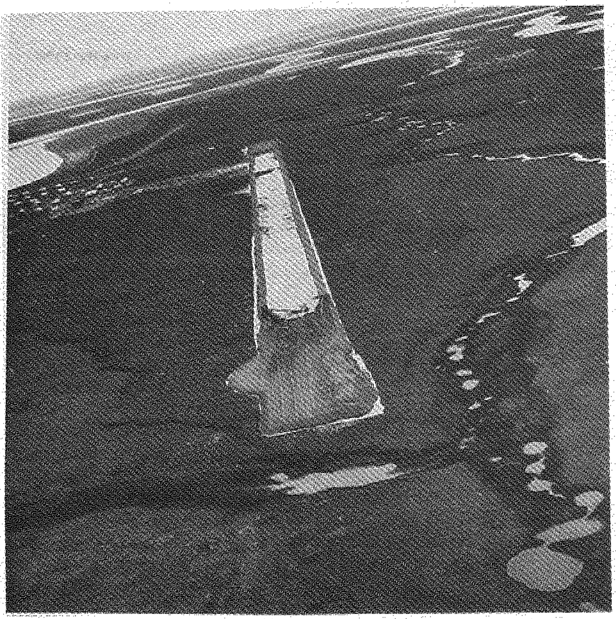


Fig. 7 Runway while under construction, using dikes to contain dredged material, at Nuiqsut, 1981.

the runway, parking apron and connecting roads was obtained by removing 10 to 15 m from the top of a small knoll. Generating sufficient quantities of select material for the surface course appears to be one of the biggest problems with rock-filled runways.

The recent development of small, transportable dredges has provided an economical method of constructing runways during the summer months, particularly for native villages along the coast and major rivers (LaVielle et al., 1983). An aerial photograph of the first such dredging operations, at Nuiqsut, is shown in Figure 7. The construction of dikes to contain the dredged material, including the use of geofabrics, is an important feature of this construction technique. Further protection of the permafrost at the shoulders of such runways is accomplished by the use of supplemental pipes to convey the excess water back to the river. Such dredge operations can also efficiently and economically generate large volumes of sands and gravels, which can be stockpiled on gravel bars and beaches for the later construction of conventional gravel or insulated gravel runways, which might be many miles from the river.

Combinations and Other Types

At some work areas the closest lake may be large enough only for a 760-m STOL runway, but a 1600-m runway may also be required. To avoid constructing such a long runway entirely on the tundra, a combination of a lake ice and snow/iced runway can be employed (Mitchell, 1981). This can only be accomplished when there are no abrupt changes in elevation along the shore of the lake. Fortunately, in some situations, there are shallows or marshy areas at one end of the lake that can be utilized, although a transitional ramp of compacted snow or cracked ice may be necessary. In other cases small peninsulas or dividing strips of land between two lakes can be bridged with snow and ice to provide the required runway length. Thus, combinations of airfield construction techniques should always be considered as practical solutions.

Landing mats, to stabilize beach gravels, have been effectively used at the Point Barrow airfield for over 40 years. The matting used was of the pierced steel plank type, used extensively during and immediately following World War 2 (Reed, 1958). While often distressed by corrosion and damaged by snow plows, it has provided excellent service, far beyond its intended life. While extruded aluminum mattings have been tested at the Inigok well site, both with and without an underlayment of insulation, to date no runways have actually been built with such mats on the North Slope (Crory, 1988).

Geofabrics and grid confinement systems are emerging alternatives for use in runways, especially when quality gravels are not available (Burns, 1979; Bell et al., 1983). Such construction materials also provide options for runway overruns, shoulders and parking aprons. They also offer new ways of stabilizing soils that otherwise would suffer from erosion, and can also be helpful in reestablishing vegetation. Other methods of constructing airfields are to be expected with the advent of new products, particularly chemical stabilizers.

RESEARCH RECOMMENDATIONS

The increasing use of computers and recent advancements in thermal modeling can significantly improve the methods of analysis used in the design of airfields on permafrost. However, the detail to which all interrelated parameters can be integrated into the computer models is rapidly approaching an impasse, in that the data input needed is just not available. Accordingly there should be much greater emphasis on the collection of air and ground temperatures at all operating airfields and at abandoned airstrips, where the snow cover is no longer being removed. Similar recommendations on the need for reliable field measurements for a proper understanding of the impact of all construction were made in the excellent review paper by Gold and Lachenbruch (1973). Thermally, the conditions of a paved or gravel-surfaced runway or of an active or inactive runway should be quite different. Contrary to the opinion that ground temperatures on the North Slope are much lower than elsewhere in Alaska, they are not so everywhere. There is a distinct warming and thinning of the permafrost as one proceeds southward, with temperatures at depth being very close to the freezing point at the base of the Brooks Range and in the nearby Foothills (Osterkamp and Payne, 1981).

Within the vast expanse of the North Slope, and elsewhere in Alaska, there are few weather stations. Still it is vital for the design of an airfield to know the wind directions and speeds, precipitation, temperature and freezing and thawing indices. Seasonal ground temperatures, taken over many years, are also essential. The extent to which the 10-year records of ground temperatures at Barrow (Aitken, 1965) have been utilized is perhaps the best proof of the value of such observations. With today's compact weather stations and small battery operated recorders, one should be capable of duplicating that 1946-1956 effort at other sites, with only annual visits to remove the data tapes and change the batteries. There is much to be learned from existing and abandoned airfields that can be transferred to the design of future airfields.

SUMMARY AND CONCLUSIONS

Airfields constructed throughout northern Alaska have provided vital transportation to both villages and other facilities. Winter airfields, on both snow and ice, have served an important function in petroleum exploration programs. Gravel surfaced runways on permafrost have performed well over the years and can be easily maintained. Virtually all abandoned gravel airstrips continue to find new uses, even if for brief periods each summer. All existing and new airfields should be instrumented to confirm design assumptions, particularly long-term thermal conditions. This field research will be increasingly important as the cost of construction of such airfields increases.

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A DEVELOPMENT OF THE ARCTIC TUNDRA IN YAMAL

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The industrial expansion is one of the vital concerns in Arctic regions, which become to be involved ever more in the sphere of the society's strategic interests. The problem of resource development and utilization is "a cross-road" of political, economic and social interests, "battle-field", where a compromise should be found between the urgent needs of the living today and concern for the future generations.

Unfortunately, our successors won't interfere with the decissions we make today, so, as a rule, among all the alternatives available the most profitable in the light of today's understanding are chosen first and foremost. Humankind is undertaking the first efforts to bring its actions in line with the remote future image.

What is going on in Tyumen North today presents the unique experience of a wide-scale intrusion into the polar regions, which remained for centuries the private domain of the northern aboriginal peoples. The experience is positive and, at the same time, negative, hence, we have come to a point, where the most serious conclusions should be drawn.

Man has demonstrated his power. Under the severe natural and climatic conditions he released and put into practice the energy, accumulated by the biosphere for milleniums, and put to use oil and gas rivers to the development of our civilization. And it is only natural that the criteria for our present-day activities as seen by future generations will be the amount of destroyed irretrievable natural resources and harm done to nature.

The situation at present has given rise to a challenging question: why is it that aboriginals of the Tyumen North began to feel fear and deep concern for the destiny of their severe but still viable land? The Nenets woman-writer ANNA NERKAGI points out: the threads of pipelines cut off the veins of the native traditional occupations - reindeer-breeding, hunting, fishing. By these words she comes to the bottom of the social-economic contradictions and emphasizes the fallacy of the traditional technologies and stereotype thinking imposed on the industrial development of the polar tundra.

Yamal, being the epicentre of unfolding contradictions, "the very edge of the Earth", beyond which the arctic desert lies, should become a prooving ground of our maturity, our readiness for reasonable management in Arctic, aimed at avoiding disturbances of the fragile landscape and destroying the system of moral values of native peoples.

By now we have encountered the ecological problem incomparable in its acuteness with no other place but Yamal. Today Yamal numberes 15 thousand people, 400 thousand reindeer, from 15 to 30 lakes per square kilometer, tens of billions of cubic meters of gas deposits, and there are the new promising prognoses of the still greater oil reserves. Untill the first geologists came to Yamal the man-nature relationship went on in harmony, spanning for centuries. As of today, we have reached a point of upcoming ecological crisis, as compared with that in the North Africa, at dawn of civilization, caused by excessive irrigation and animals' pasturing. In Yamal it is the trenches of bulldozers, not reindeer herds, that may lead up to irreversible ecological process. Railway roads are still being planned, but 6 thousand hectare of reindeer pasture have already fallen a victim to the increased pace of northern development.

The Bovanenkovo deposit alone, which will become part of the Soviet Central Gas Supply System in a few years, numberes 12 animal species and over 50 bird species, 3 of which are included in the Read Book. The territory around the deposit is a reservation for reproduction of some species of wildlife, insuring maintainance of the optimum density of living beings after they are being regularly diminished.

The problem of atmospheric pollution, especially by nitric oxides, is of immediate importance. Even if the nitric emissions are harmless to people, they are disastrous to the surface lichen and moss cover. And what makes things even worse in Yamal is the low ability of soils for self-purifications and frequent (up to 200 days a year) fogs, making nitric substances more toxic.

The common task for all Arctic states is to get away with a fetish of industrialization, to have a new look at the last pure, as regards the ecology, spaces beyond the North Polar.

This is of particular importance for Arctic, as it is becoming the sphere of international scientific cooperation, natural laboratory and the source of new natural resources.

Let us now consider the possible sources of the ecological catastrophe in Yamal. One of them is associated, first and foremost, with the low stability of natural complexes, even in the absence of human activities. The high activity of cryogenic processes, mainly thermal erosion, thermokarst and solifluction, determines spontaneous instability of the slightly declined terrain. Some slopes are slumping continuously, others being stable for many years, are subsiding in what seems to be an abrupt manner, exposing ice masses. Appearance and development of ravines occurs, as well as lake formation through enlargement of perched ponds. Thermal erosion and thawing result in subsequent coastline retrieval by 0,5-3 m a year. The lakes are enlarging, but then often become partially or completely drained. The relief of Yamal, determined by the natural bifurcation mechanism, presents an amazing variety. The same bifurcation mechanism, as far as relief-formation is concerned, are associated as well with human-induced activities. Trails made by off-road vehicular traffic spread out in all directions, for one and the same road cannot receive many vehicular passes, and even man made roads cause bifurcation.

The ecosystem of Yamal has encountered arising micro- and macrocatastrophes, found in no other place but here. New systems are born spontaneously, which are likely to change in quantity and, furthermore, in quality, thus giving rise to a new irreversible process.

Demography can be regarded as another source of ecological disastor.

The discovery of oil and gas deposits in Western Siberia led to the population rise by 1,5 million people in the Tyumen North. New towns were built with the population up to 100 thousand and over. Town-building in Western Siberia is conditioned above all by the structure of underground petroleum resources.

The trend for building "utilitarian" towns, oriented as they were towards the urgent needs, triggered a number of increasing social, economic and ecological problems. The above-mentioned bifurcation mechanism led up to irreversible social phenomenon - self-creation and self-development of towns. In many towns a comparatively small part of population is needed for realization of initial goals, the majority require new production field be formed.

The unmanagable growth of production forces will be disastrous under the natural-climatic conditions of Yamal. In case the production forces are put under control, but don't agree with the environment adaptation ability and our possibilities to make amends for the losses, the result will be the same. The Yamal tundra, for example, can't already provide 400-thousand reindeer herd with feeding, for the pasture revegetation ability covers about 300 thousand reindeer. And the Bovanenkovo deposit alone is found to remove the surface cover, able to support 30 thousand reindeer. Evidently, that the main difficulties in the polar tundra development are associated with the cryolithozone peculiarities in Yamal. Permafrost occurs in all the elements of the surface. Continuous permafrost with rare taliks occurs below large lakes and rivers. The observed maximum thickness of permafrost layer exceeds 450 m and the minimum ground temperature is minus -8°C. Permafrost is characterized by deposits with high ice content, contains wedge ice and sheets of ice of the size 20x100x200 m and more. The major part of permafrost is salted up to great depths, which impacts the underground water supply and makes the problem of pure water protection more complicated. Permafrost sites of high ice content and of thin protective cover is the main latent source of irreversible process in Yamal ecosystem.

Yamal itself presents the incarnation of four-dementional supersensitivity to the impact of everchanging conditions.

It is known that the increase of the average temperature on the Earth's surface is accompanied by the temperature drop between the equator and the Pole. Nevertheless, according to the long-term observations, the mean temperature in the equator is practically constant. All the temperature changes are associated with polar zones. A slight decrease in the mean temperature on the Earth (by 0,5 cause the temperature decrease in the North by 2-3°C with all corresponding consequences.

The thin thermal protective cover makes the permafrost sensitive to any energetic influences. Deprived of its protective ability, it may give rise to activization of processes to great depths with unpredictable results.

SUMMARY

Supersensitivity of the arctic tundra to the matter and energy fluxes requires mostly scientific instead of energy consuming technologies.

The traditional task of ecology is to maintain the ecosystem ability for self-rehabilitation. But the peculiarities of vegetation cover and of permafrost itself make this task practically unreal, and contradictory to natural processes. Hence, the main task of ecology is to provide the smooth transition of one ecosystem into another. The new-formed ecosystem should possess screan and biological properties, able to minimize the transitional losses.

Human activities impact on the global processes, disturbing existing landforms, increases possibility of wide-scale ecological disastors in Arctic.

Ecosystems of Yamal-type disappear, naturally or anthropogenically, under the water layer, caused by excessive industrialization or by the significant warming up of the Earth's atmosphere. Inevitable changes in circulation processes in the atmosphere and ocean will trigger new process in the biosphere. This makes evident the necessity of sound investigations on the tendencies of anthropogenic impact. And it is up to geocryologists to contribute to modeling of global processes and their evolution, with special reference to cryosphere.

ADDITIONAL PAPER:

PERIGLACIAL FORMS RELATED TO TERRAIN PARAMETERS IN JOTUNHEIMEN, SOUTHERN NORWAY

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SYNOPSIS An area of 19.2 km^2 located around Juvvasshytta in northern Jotunheimen, central southern Norway, is mapped in detail with emphasis on periglacial features and their relation to to terrain parameters. Periglacial forms are the most significant geomorphological features of the area. The areal cover of patterned ground is from 20% to 50% in a continuous cover of till above 1750 m a.s.l. In areas above 2000 m a.s.l. the areal cover is generally less than 10% due to a smaller amount of debris. These values are calculated from slope gradients less than 10 degrees. Sorted steps and stripes have a well-defined lower limit slope gradients at two degrees and three degrees. The upper limit is not well-defined. These distributions have a tail at steeper gradients. Block tongues have a normal distribution with the average at 13 degrees.
Alpine permafrost is well-documented at elevations higher than 1700 m a.s.l. There is a positive correlation between the existence of periglacial forms and the presence of permafrost.

GEOMORPHOLOGICAL MAPPING

The geomorphological mapping is based on high-quality vertical aerial photographs on the scale 1 : 10 000. The smallest interpretable objects have a diameter of 0.2m - 0.3m. Construction of the contour map on the scale 1 : 10 000 and interpretation of geomorphological features have been conducted with a stereo instrument (Stereo-simplex G 6) at the Laboratory of Remote Sensing and Thematic Mapping, University of Oslo. A digital terrain model was developed from the analog map to investigate the relations between periglacial features on one side and slope gradient and exposure on the other.

Results from this mapping show that debris is well represented at different elevations and different exposures. Seventy-four percent of the total investigated area has a continuous cover of unconsolidated material. This make the area well-suited for a case study of zonation of periglacial forms. The area is 55% block covered, mostly frost-sorted till and weathered in situ deposits. Thirteen percent is covered by glaciers, 12% by patterned ground, 7% by unsorted till, 7% by snow patches surviving the thaw and 3% by lake surface, which is at 1836 m a.s.l.

The mapping of patterned ground is based on Washburns classification (Washburn, 1956). Sorted forms in flat terrain have irregular patterns. These areas have been classified as sorted forms in flat terrain, which means sorted polygons, nets and circles. Sorted steps are forms elongated in the direction of the steepest slope. The centres of fine domains are usually flat with a marked step of coarser material at the border. Sorted stripes are stripes of coarser material in a matrix of debris. Block tongues have a border of coarser material, usually with a centre of fine domains with the same gradient as the coarser material at the border of the tongue.

The boundary between mainly weathered in situ deposits and frost sorted till is at 1950 m a.s.l. In areas with a slope gradient of more than 8-10 degrees, this boundary is lower, probably due to solifluction and block creep. Grain size analyses show that the centres of fine domains are made up of 30%-40% material finer than 0.125 mm, with no significant differences between till and weathered in situ deposits. The latter, however, have a reduced total amount of debris.

The bedrock of the area consists mainly of gabbro with intrusions of peridotite. This lack of homogeniety in the bedrock does noe affect the zonation of periglacial forms.

ALPINE PERMAFROST

The mapped area represents altitudes from 1550 m a.s.l. to 2200 m a.s.l. The estimated mean annual air temperature (MAAT) is -2.6°C at 1500 m a.s.l. and -6.4°C at 2200 m a.s.l. This estimate is based on interpolation and extra-

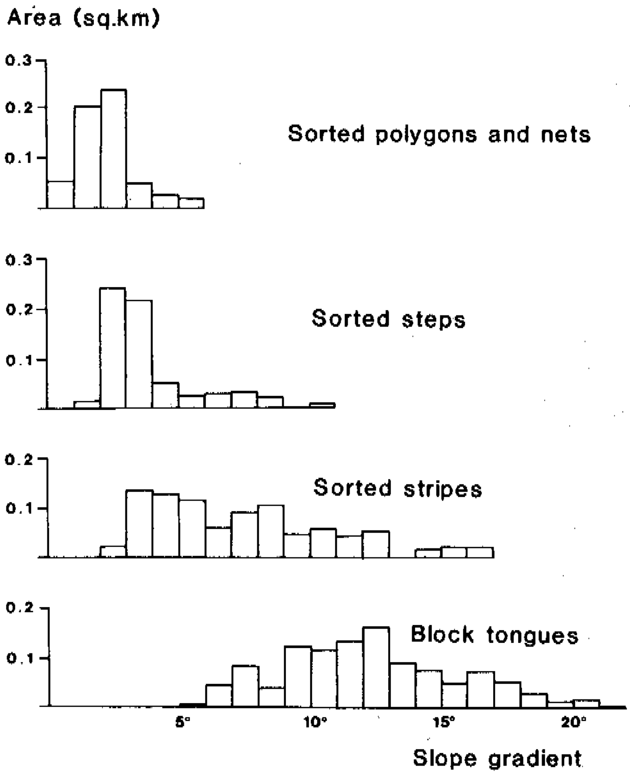

Fig.1 Relation between slope gradient and periglacial form within the mapped area.

polation of MAAT from 11 climate stations in central southern Norway. Mean annual ground temperature (MAGT) has been measured as -2.1 °C at 1850 m a.s.l. close to Juvvasshytta, which is 2.4 °C higher than the estimated MAAT (Ødegård, 1986, Liestøl, 1986). This MAGT is higher than estimations made bye King (1984).

The thickness of the permafrost is probably 50-100 m. There are no measurements of the geothermal gradient in the area. The site where the temperature measurements have been conducted generally has a snowcover of 0-0.15m from September to April. Maximum snow cover has been measured as 0.5m in late May. In flat areas and on slopes exposed to the south and west, the snowcover melts during June. MAGT at 1850 m a.s.l. indicates a lower limit of alpine permafrost in flat areas at 1500-1700 m a.s.l. This lower limit of alpine permafrost corresponds well with the lower limit of ice-cored morains in Jotunheimen at 1550 m a.s.l. (Østrem, 1964). Patchy discontinuous permafrost has been reported between 1450 m a.s.l. and 1700 m a.s.l. in the Leirvassbu area (King, 1984)

SLOPE GRADIENT AND FORM

Figure 1 illustrates the relation between slope gradient and periglacial forms within the mapped area. The investigation is based on grid of points at 80 meters separation. The whole area of investigation is 19.2 km^2. Of this area, 2.3 km^2 is made up of sorted forms with centres of fine domains (341 gridpoints) and 10.7 km^2 (1672 gridpoints) block covered areas, including block tongues. The total number of block tongues is 522. The number of gridpoints with occurances of block tongues is 169 which corresponds to an area of 1.2 km^2.

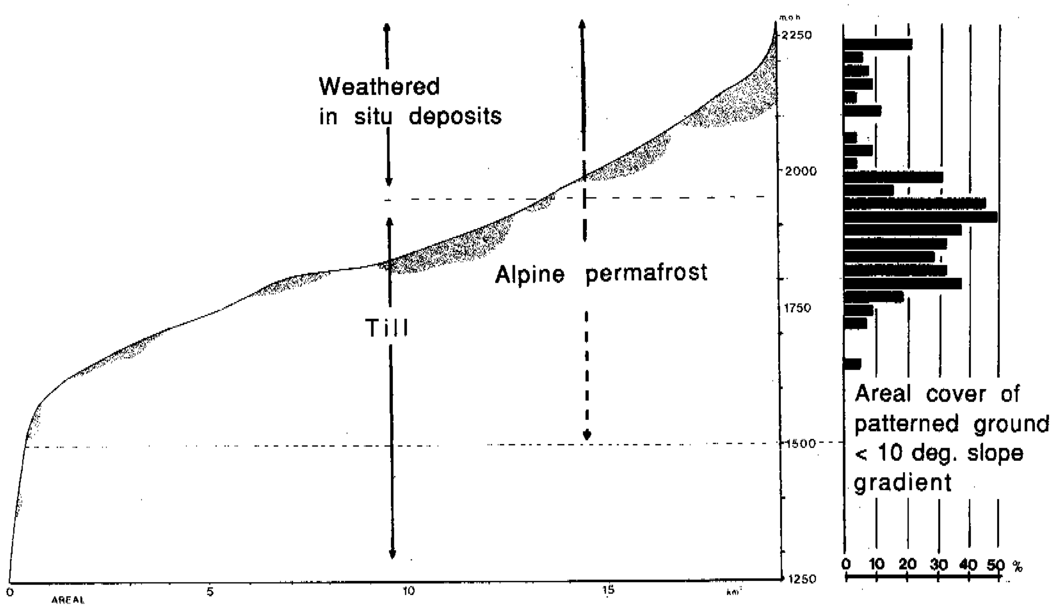

Fig.2 The distribution of periglacial forms in relation to height. The left half of the figure is a hypsographic curve of the area, the right half shows the areal cover of patterned ground.

Sorted steps and stripes have a sharp lower limit with a tail at steeper gradients. Due to the resolution of the gradient measurements (80m x 80m), there are very few gradients lower than one degree. The lower limit for sorted steps is close to 2.0 degrees and the lower limit for sorted stripes is at about 3.0 degrees. Block tongues have a normal distribution, with the average at 13 degrees. Table 1 shows typical gradients for different geomorphological forms.

ALTITUDE AND FORM

In the investigated area the areal cover of patterned ground generally increases with height above 1750 m a.s.l. At 1950 m a.s.l. there is discontinuity due to a change in the amount of debris. Block fields above this altitude are dominated by in situ weathered deposits, but erratics from local bedrock exist all over the area. Below, a continuous cover of frost-sorted till dominates. The origin of this change in morphology is probably from the last deglaciation.

The distribution of periglacial forms in relation to elevation is illustrated in Fig.2. Well-devloped sorted forms correspond well with the assumed distribution of alpine permafrost. Over 20% of areas assumed to have alpine permafrost are cover with patterned ground. In non-permafrost areas the areal cover is less than 1%. This estimate is based on slope gradients less than 10 degrees.

DISCUSSION

The distribution of fossil versus active forms in the area is an important aspect concerning the present zonation. At high elevations above 1900 m a.s.l. there are active forms at present. The size of the centres of fines is from 0.5m - 1.5 m. The debth of the active layer is from 0.5m - 0.8m.

From 1700-1900 m a.s.l. the fine domains have a non-vegetated centre which indicates active frostprocesses. There is no activity at the surface of the borders or in block covered areas, which are heavily lichen covered. At lower elevation there are few indications of active frost sorting. In the height interval 1700-1900 m a.s.l. the size of the centres of fines is typically 3m - 5m. The debth of the active layer is 1.5m - 2.0 m.

TABLE 1

Slope gradients for different features in the area.

	Average	Number
Sorted polygons and nets	2.0	90
Sorted steps	3.9	101
Sorted stripes	7.3	147
Block tongues	12.7	169

From the size of the features related to the present depth of the active layer, there is reason to believe that the features are in equilibrium with the present climatic conditions. The fossil appearance is due to an equilibrium in the soil/block moving processes at the borders, probably because the total amount of fines is not sufficient to maintain active frost processes in the inter circular areas. The block size in the inter circular areas are typically 0.3m - 1.0m.

Key factors determining the zonation of periglacial forms are the water content of the soil, the total amount of fines and slope gradient. Because of large areas with alpine permafrost at high elevations, the water content is generally increasing with height. West-facing slopes have few snow patches in mid-summer. Here, the water supply during summer is less than that of east-facing slopes, where numerous snowpatches survive the thaw.

Patterned ground is most frequent on east-facing slopes in the interval 1800-1900 m a.s.l., which has a 35% areal cover of patterned ground on slope gradients less than 10 degrees. Areas above 1900 m a.s.l. have scattered sorted forms with diameter seldom larger than 1.0 m. In situ weathered deposits dominate. From these results the general zonation of periglacial features might be explained bye the soil's water content, the amount of fines in the soil and the slope gradient. The presence of alpine permafrost is a critical factor affecting the water content. Height intervals with an areal cover of patterned ground more than 5% - 10% are in the alpine permafrost zone. There is reason to believe that other similar areas in southern Norway are in alpine permafrost.

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GROUND ICE REGIONS OF NORTH AMERICA

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SYNOPSIS Six major ground ice regions are recognized within the permafrost zones of North America. They are distinguished on the basis of the form of occurrence and extent of ground ice. The regions have distinctive suites of bedrock geology, Quaternary history, surficial geology, hydrogeology and climatic history. The six regions and key distinguishing characteristics are:

1. Nothern Alaska: unglaciated, marine transgression/regression sequence, eolian activity; permafrost continuous, ground ice mainly as wedges and segregation ice, pingos locally common.
2. Tuktoyaktuk Coastlands, Banks Island, Victoria Island: glacial margin, thick surficial deposits; permafrost continuous, extensive massive ground ice, wedges common, pingos locally very common.
3. Queen Elizabeth Islands: partially(?) glaciated, weakly lithified bedrock, variable distribution of surficial deposits; permafrost continuous, massive ground ice and ice wedges locally(?) significant, pingos rare.
4. Baffin-Keewatin-Ungava: heavily glaciated, indurated bedrock, limited surficial deposits; permafrost continuous to discontinuous, extent of ground ice limited, pingos very rare.
5. Mackenzie Valley-Northern Provinces: heavily glaciated, thick surficial deposits, extensive peatlands; permafrost continuous to discontinuous, ground ice widespread in peatlands and lacustrine deposits, pingos rare, palsas common.
6. Cordillera: degree of glaciation variable, bedrock and topography complex; permafrost discontinuous and alpine, ground ice conditions complex, open system (hydraulic) pingos common in northern part. This region may be subdivided into three:
   (a) Central and Southern Alaska and Interior Yukon: partially glaciated;
   (b) Southern Yukon and British Columbia: glaciated; and
   (c) Western United States: very limited glaciation.

In addition, two azonal situations are recognized: (i) areas of saline soils, and (ii) areas of subsea permafrost, both with limited ground ice.

A GROUND ICE MAP OF NORTHWESTERN CANADA

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SYNOPSIS A new permafrost map of northwestern Canada, emphasizing ground ice conditions is presented. The map, at a scale of 1:1 000 000, covers the Yukon, the lower Mackenzie Valley and Delta area, and the Tuktoyaktuk Coastlands area of the western Arctic. The map was compiled from existing maps of surficial geology and terrain conditions, supplemented by site-specific observations. The extent of permafrost within each map unit is classified in a 6-step scale, from continuous (1 class), through discontinuous (4 classes) to absent (1 class). Ground ice content is classified in an 8-step scale, from high, through moderate and low, to nil, with intergrades and a variable class. The form in which the ground ice occurs is described in the legend, along with pertinent information on its geological and geomorphological setting. Known locations of selected geocryological features, such as pingos, rock glaciers and ice-cored moraines are shown by symbols. The gross distribution of permafrost and ground ice in this region is controlled by latitude, elevation, distance from the Pacific and Arctic Oceans, the seasonal positions of major air masses and glacial history. On a more local scale, geological, geomorphological, topographic and vegetational controls are more significant. In all areas, human activity can no longer be ignored as a factor influencing permafrost stability.

REGIONAL FACTORS OF PERMAFROST DISTRIBUTION AND THICKNESS, HUDSON BAY COAST, QUÉBEC, CANADA

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(Paper in Vol. 1 p. 199)

ROCK GLACIERS IN NORWAY, SVALBARD INCLUDED

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Mapping of rock glaciers, mainly by means of aerial photographs is carried out in Norway, Svalbard included. Sporadic mapping of rock glaciers was done at the Department of Geography in the 70's. Of more importance, however, is a glacial geological map of Norway (Sollid & Torp 1984) and a map of glacial geology and geomorphology of Svalbard (Kristiansen & Sollid 1986). Both maps belong to the Norwegian National Atlas. On the map of Norway mainland, ice-marginal moraines deposited by local glaciers/rock glacier deposits are presented in one group, but they were originally mapped separately. Most important, are the maps of Nordre Andøya (Flakstad et al.) and Kåfjord (Tolgensbakk & Sollid 1988). In these areas, as well as in selected sites in Svalbard, a systematic study of rock glaciers has been undertaken by the authors. Prelimenary results have been presented by Sollid (1987) and Sollid & Tolgensbakk (1988).

Rock glaciers are of permafrost origin. In Svalbard, which has thick, continuous permafrost in areas not covered by glaciers, most rock glaciers are still active. This is probably less common for rock glaciers in mainland Norway. In Svalbard most rock glaciers are found in western Spitsbergen which was deglaciated early. On the mainland rock glaciers are mostly located outside the Younger Dryas (10-11 ky. B.P.) end moraine system, indicating that the unglaciated areas had permafrost at that time. Alpine permafrost exists in the mainland today, probably also in some of the rock glaciers. The greatest frequency of rock glaciers on the mainland is in the central parts of Troms, northern Norway, where the rock glaciers are often located fairly close to post-glacial fault lines. Rock slides might here have played a major role concerning the origin of the rock glaciers. In some places the rock glaciers mainly consists of very big boulders.

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DETAILED MAPPING OF PERIGLACIAL FORMS IN JOTUNHEIMEN, SOUTHERN NORWAY

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(Paper in Vol. 3 p. 59)

FROST HEAVE

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(Paper in Vol. 1 p. 344)

A MATHEMATICAL MODEL OF FROST HEAVE IN GRANULAR MATERIALS

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(Paper in Vol. 1 p. 370)

MICROSTRUCTURE OF FROZEN SOILS EXAMINED BY SEM

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(Paper in Vol. 1 p. 390)

SPATIAL VARIATION IN SEASONAL FROST HEAVE CYCLES

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(Paper in Vol. 1 p. 436)

THE FORMATION OF PEDOGENIC CARBONATES ON SVALBARD: THE INFLUENCE OF COLD TEMPERATURES AND FREEZING

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(Paper in Vol. 1 p. 467)

GENESIS OF ARCTIC BROWN SOILS (PERGELIC CRYOCHREPT) IN SVALBARD

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(Paper in Vol. 1 p. 478)

CHEMICAL WEATHERING IN PERMAFROST REGIONS OF ANTARCTICA: GREAT WALL STATION OF CHINA, CASEY STATION AND DAVIS STATION OF AUSTRALIA

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(Paper in Vol. 1 p. 511)

SOIL DISPLACEMENT IN SORTED CIRCLES, RESOLUTE AREA, CORNWALLIS ISLAND, CANADIAN HIGH ARCTIC

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SYNOPSIS Lines of dowels originally inserted to depths of 100 and 200 mm across sorted circles were repeatedly surveyed over a 3-year period. Measurements of dowel heave, lean, and lateral displacement provided evidence on circle activity. Subject to certain assumptions, the evidence suggests that the cumulative soil displacement was radial and that its velocity near the surface increased linearly from zero at the center of a circle to a maximum at the stony border. This permits an estimate of the minimum age of circles. Radial displacement decreased with depth and was probably negligible below 100 mm. Although the radial displacement pattern supports the convection hypothesis for the development of sorted circles, it does not necessarily prove their origin.

PERENNIAL DISCHARGE OF SUBPERMAFROST GROUNDWATER IN TWO SMALL DRAINAGE BASINS, YUKON, CANADA

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(Paper in Vol. 1 p. 639)

A FIRST APPROACH TO THE SYSTEMATIC STUDY OF THE ROCK GLACIERS IN THE ITALIAN ALPS

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(Paper in Vol. 1 p. 712)

THERMOKARST AND CANOE-SHAPED BLOWOUT DUNES IN A SUBARCTIC DUNE FIELD, KOBUK VALLEY, NORTHWESTERN ALASKA, U.S.A.

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SYNOPSIS During a reconnaissance geomorphological survey of late-Wisconsinan to partly still active dune fields in the Central Kobuk Valley a number of canoe-shaped blowout dunes were encountered, mainly stabilized by spruce and birch forests. This part of the floodplain of the Kobuk River consists of possibly ice-rich, silty and sandy, fluvial and eolian sediments and lies at the transition of discontinuous to continuous permafrost. Simple and complex blowouts consist of elongated, flat-bottomed, up to 10-15 m deep deflation depressions sharply outlined by narrow longitudinal or arcuate ridges and parabolic dune heads at the western, eastern or both ends of the blowouts.

These dune forms differ from elongate parabolic dunes, hairpin dunes or winddrift dunes in that the arcuate ridges sometimes converge at both ends of the blowouts. The forms are extremely uniform oriented (WNW-ESE) and vary in length from 0.5-6 km. On the border ridges as well as in the deflation depressions complex systems of partly reactivated, secondary parabolic and sinuous transverse ridges have developed. The blowouts are dissected by fluvial erosion and exhibit modification by thermokarst. In the depressions occasionally wet peaty areas and elongate ponds occur.

It is suggested that this dune type originates by initiation of spot blowouts leading to locally accelerated wind velocities and greater bed scour controlled by a high velocity opposite bimodal wind regime. Local permafrost degradation may be caused by development of the blowouts in connection with local disruption of the vegetation. Concentration of water in summer and of snow in winter in the initial deflation basins could promote further melting of ground ice, thereby inducing so-called self-developing thermokarst. Gradual wasting of the dunes will ultimately result in stabilization by vegetation at the rims.

MICROTOPOGRAPHIC THERMAL CONTRASTS, NORTHERN ALASKA

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(Paper in Vol. 1 p. 819)

DETERMINATION OF EXTENT AND STIFFNESS OF FROZEN GROUND USING SEISMIC SURFACE WAVES

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SYNOPSIS The poster presentation discusses a feasibility research study on an innovative new seismic technique, the Spectral Analysis of Surface Waves (SASW) method, for locating frozen and unfrozen ground rapidly and accurately through measurement of the surface wave velocity profile. The velocity of seismic waves is much faster in frozen ground than unfrozen ground. The SASW method has already been used to predict modulus and thickness profiles for pavements and for investigations of soil sites, dams, and fills.

The presentation includes an explanation of the methodology and principals of the SASW survey, an example dispersion curve (wavelength versus phase velocity), and seismic shear wave velocity and modulus profiles from three field sites in the Fairbanks, Alaska area (the University of Alaska at Fairbanks hayfield, the Loftus School and the CRREL Farmers Loop Facility). Comparisons are presented of results of downhole seismic shear wave velocity tests and soil borings for comparison of frozen/thawed profile results. The feasibility results show excellent agreement between surface wave and downhole seismic velocity results and clearly indicate the increased stiffness associated with the permafrost.

One of the most exciting aspects of the method is its capability to detect low velocity layers underlying higher velocity layers with measurements made on the ground surface. This capability of the SASW method overcomes the traditional limitation on the use of the seismic refraction method. Thus, the SASW method shows strong potential for locating frozen and unfrozen ground layers even when the ground surface is frozen and the underlying soils are thawed. The potential benefits of accurate knowledge of frozen ground conditions to engineers, contractors and the public are tremendous for improved performance and economy of construction and maintenance of roads, buildings, runways, and pipelines.

This Phase I research study was sponsored by the Small Business Innovation Research Program of the National Science Foundation through their Polar Operations Division and conducted by Olson-Church, Inc.

ORIGIN OF MASSIVE GROUND ICE ON TUKTOYAKTUK PENINSULA, NORTHWEST TERRITORIES, CANADA: A REVIEW OF STRATIGRAPHIC AND GEOMORPHIC EVIDENCE

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(Paper in Vol. 1 p. 850)

PERIGLACIAL GEOMORPHOLOGY OF KVADEHUKSLETTA, SVALBARD

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SYNOPSIS The map Kvadehuksletta kvartærgeologi og geomorfologi made by Tolgensbakk & Sollid (1987) will be demonstrated. The base map which is constructed by the Norwegian Polar Research Institute (NP), is in scale 1:10 000 with contour interval of 10 m. The authors have supplemented this base map with a 5 m contour interval. The area mapped in detail concerns Quaternary geology and periglacial features. The legend of the map has 73 different thematic symbols, among them 13 areal ones.

The Kvadehuksletta area is situated on the southern flank of Kongsfjorden, Spitsbergen, is manteled by marine raised shore beaches at Brøggerhalvøya's headland. The shore-line displacement has been studied by Forman et al. (1987). The area has well developed periglacial patterned ground features, especially sorted circles. At selected sites, circles have recently been studied by Dr. B. Hallet, QRC, University of Washington (Hallet & Prestrud 1986) in a cooperation project with the authors. The authors are also studying rock glaciers, some of them located within the map area. Measurements indicate that these glaciers are active. Rock glaciers are very common in western Spitsbergen as shown by Kristiansen & Sollid (1986).

An interesting part of the Kvadehuksletta study is the investigation of the genesis of the fines and the distribution of the weathering material. Due to cold-based ice conditions during glaciation periods the surface might have been little disturbed through a long period of time.

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FLUVIO-AEOLIAN INTERACTION IN A REGION OF CONTINUOUS PERMAFROST

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(Paper in Vol. 1 p. 876)

OBSERVATIONS OF SORTED CIRCLE ACTIVITY, CENTRAL ALASKA

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(Paper in Vol. 1 p. 886)

PLEISTOCENE PERIGLACIAL CONDITIONS, U.S. MIDLANDS

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SYNOPSIS During the past decade, ice-wedge casts and patterned ground in southwestern North Dakota, northeastern Nebraska, and Illinois have added to the previous observations in southeastern South Dakota, eastern Iowa, and central Indiana to give a picture of a Wisconsinan ice marginal permafrost belt running from Indiana to eastern Montana, increasing westward in width, sporadic in the east to nearly continuous in the west. Mean annual temperatures along the ice margin probably ranged from about -4^{0} C in Indiana to -12^{0} C or lower across Nebraska and the Dakotas, where they were accompanied by strong northwesterly winds. Interpretations based largely on faunal assemblages of vertebrates and mollusks that had lived 200 km or more from the ice margin have suggested that the ice-marginal environment during pre-Wisconsinan glaciations may have been less severe than those of the last glaciation. Pollen records are sparse, and only limited physical evidence has been reported.

In Indiana, ice-marginal silts contain 3 species typical of the subarctic (<u>Vertigo alpestris oughtoni, V. modesta, and Columella alticola</u>) and none of the warmth-loving species of the area today. In Nebraska ice marginal faunules have not been found; the nearest, in loess that post-dates the ice wedge development, also contains three alpine species (<u>Oreohelix strigosa berryi, V. modesta, C. alticola</u>). In both areas, southward from the permafrost belt, one (<u>V. alpestris</u>, Indiana, and <u>O. strigosa</u>, Nebraska) disappears almost immediately, and only <u>C. alticola</u> is present A few of the thermal contraction features in North Dakota are clearly pre-Wisconsinan, but their age, beyond that, is not known. In eastern Nebraska, earlier Pleistocene proglacial faunal assemblages suggest a climate much like that of eastern North Dakota today, as do Peoria Loess assemblages from the same area and southward. Comparable similarities can be found among vertebrate faunal assemblages from northern Nebraska to southwestern Kansas. Ice-marginal conditions probably did not differ greatly in the Central Lowlands and Great Plains during the pre-Wisconsinan glaciations from those that developed during Wisconsinan time.

LANDSLIDE MOTION IN DISCONTINUOUS PERMAFROST

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(Paper in Vol. 1 p. 897)

PERMAFROST AND TERRAIN PRELIMINARY MONITORING RESULTS, NORMAN WELLS PIPELINE, CANADA

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(Paper in Vol. 2 p. 916)

DEVELOPMENT AND THAWING OF ICE-RICH PERMAFROST AROUND CHILLED PIPELINES MONITORED BY RESISTANCE GAUGES

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(Paper in Vol. 2 p. 1004)

PREDICTION OF PERMAFROST THICKNESS BY THE "TWO POINT" METHOD

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(Paper in Vol. 2 p. 965)

DISCONTINUOUS PERMAFROST MAPPING USING THE EM-31

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(Paper in Vol. 2 p. 1018)

RELATIONSHIPS BETWEEN SAMPLE QUALITY AND LABORATORY TESTING OF PERMAFROST SOILS

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SYNOPSIS The Permafrost Subcommittee of the Associate Committee on Geotechnical Research, National Research Council of Canada, is preparing a laboratory manual for testing permafrost soils. The manual will cover laboratory testing procedures to determine physical, mechanical and thermal properties. In participating in the preparation of the manual, it became apparent that poor quality samples can ruin the most sophisticated testing program. This does not imply that all testing procedures require a high quality of sample. The relationship between sample quality and laboratory testing will be presented in a tabular form. In undertaking a testing program, it is important to keep in mind the specific aim of the investigation. Once the required test methods have been chosen, then the sample quality must be given careful consideration. The quality of the sample is dependent on the type of material sampled, the in situ condition at the time of sampling, and the methods of sampling, transportation and storage.

SHALLOW FOUNDATIONS, HEAT PUMPS, ACTIVE REFRIGERATION

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SYNOPSIS Since 1986 the National Research Council has been considering the possible applications of ground source heat pump technology to Arctic problems in collaboration with the Government of the Yukon two field projects have been undertaken to evaluate the technology for slab foundations located on permafrost. Using a system of heat collector pipes located below the insulation the heat normally escaping to the ground is returned to the living space via the heat pump. This arrangement makes it possible to maintain frozen conditions in the foundation pad and sub-grade while producing a significant quantity of heat for the building itself. For many communities in northern Canada the relative costs of fuel oil and electricity are such that operating costs may be largely or entirely offset by the saving in fuel costs. Of potentially greater benefit is the reduction in initial construction costs compared with current alternatives.

FOUNDATION DESIGN ON PERMAFROST WITH HEAT PUMPS

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SYNOPSIS Bjarne Instanes A/S, Consulting Engineers, have designed more than 200 projects on Spitsbergen during the last 20 years. On the Norwegian mainland, the company has participated in several Heat Pump projects. Several years ago, the author developed the idea to use heat pumps for foundation design on Svalbard.

The idea is simple: Heat pumps use the subsoil underneath the building as energy resource. Thus the temperature underneath the foundations is lowered to approximately $-10^{0}C$. The gained energy is used for heating the building. In other words, the heat loss through the insulated floor is transported back inside the building by means of heat pumps.

The advantages of the foundation principle is:

- Creating low temperatures under foundations and thus high bearing capacity of the soils.
- Contribute to the heating of the building with low energy input.
- Economical foundation design.

Foundation design with artificial freezing of the ground by means of heat pumps, was used for the first time in Spitsbergen for a Storage Building in Sveagruva. The building has a floor area of 900 m^2. The foundation design can briefly be described as follows:

- The existing soil (characterized as Svea Clay), was leveled to minimum 0,5 m below the level of pipes for cooling fluid.
- A minimum 0,5 m thick gravel pad was laid out and compacted.
- On the gravel pad, the 40 mm cooling pipes were laid out in 1,0 m distances.
- The pipes were then embedded in a 100 mm thick concrete slab, generally 200 mm expanded polystyrene was used. In order to sustain the concentrated loads under single column footings, special quality insulation was applied.
- The space between the insulation and the floor slab is filled with a 700 mm thick layer of fine sand. The thickness of 700 mm was decided mainly to satisfy minimum slope for sewage pipes etc.
- The floor is a 150 mm thick reinforced concrete slab.
- A computer programme was developed to calculate the necessary effect, the quantity of cooling fluid, the dimensions of the cooling fluid pipes, the insulation needed etc. to obtain the design temperature in the soil.

  Necessary effect to meet the temperature required at the level of the cooling pipes ($-10^{0}C$) was calculated to approx. 16 kW. Mostly due to security reasons, there is installed two separate systems, each of 18 kW. Pumps, compressors etc. are automatically controlled by termistors placed strategically in the ground.

The system has been in operation since its completion in the autumn 1986. It has been working satisfactory without any problems, and has proved to be an economical method for foundation design on permafrost. The principle may also prove to be feasible for buildings on difficult soil conditions in areas without permafrost.

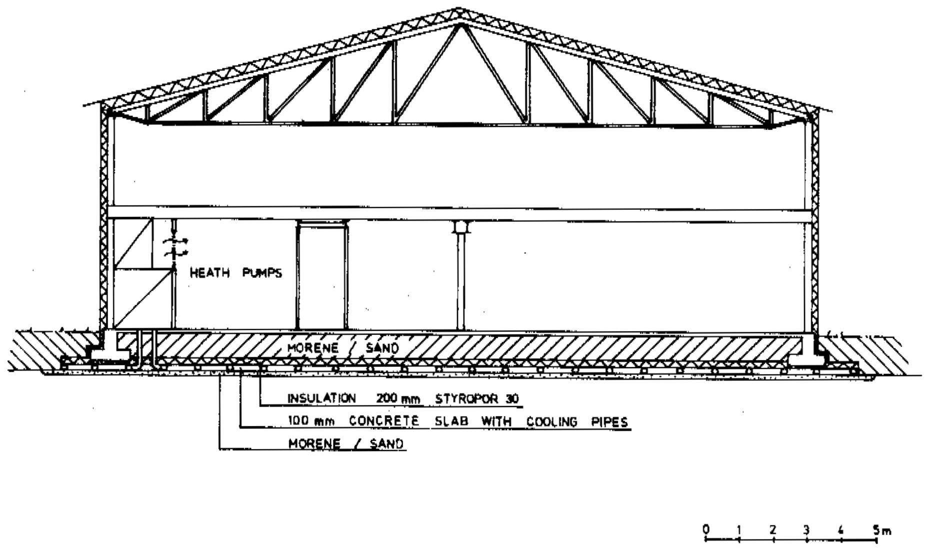

Fig. No. 1 STORAGE BUILDING, SVEAGRUVEN, 1986

BJARNE INSTANES

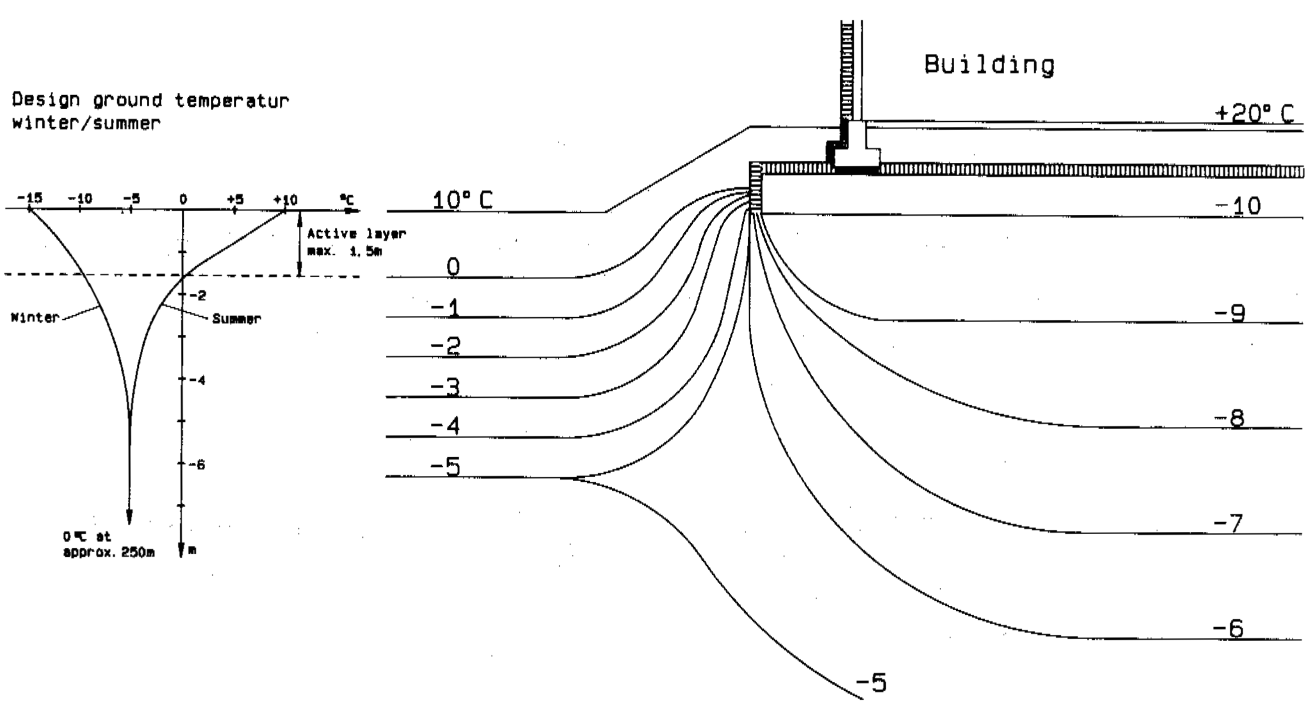

Fig. No. 2 CALCULATED SUMMER ISOTHERMES

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THAW STABILIZATION OF ROADWAY EMBANKMENTS

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(Paper in Vol. 2 p. 1352)

THE CRACKING AND ITS PREVENTING OF THE ROAD UNDER LOW TEMPERATURE

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SYNOPSIS Based on field investigations on an asphalt-paved highway in Heilongjiang Province, North-east China, features of the cold cracking of the road pavement were described, and its causes were analyzed as well. After intensively analyzing its influencing factors, measures for preventing the cold cracking were presented from various aspects including selection of pavement material, design of road structure and control of construction quality, etc.

THE RESISTANCE TO FROST HEAVE OF VARIOUS CONCRETE CANAL LINING

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(Paper in Vol. 2 p. 1482)

AIRPORT NETWORK AND HOUSING CONSTRUCTION PROGRAMMES IN NORTHERN QUEBEC, CANADA

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(Paper in Vol. 2 p. 1500)

RIVER CHANNEL DREDGING IN THE PERMAFROST-DOMINATED NORTH SLOPE

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SYNOPSIS In 1980, the North Slope Borough (NSB) began an extensive Capital Improvement Program (CIP). Most of the CIP projects required vast amounts of gravel and, because upland sources were limited, river, lake, and lagoon dredging was initiated. In many cases material was placed directly on the tundra surface where needed minimizing subaerial modification. Subaqueous change is being monitored by a continuing program initiated by the NSB.

STUDIES ON THE PLASTIC-FILM-ENCLOSED FOUNDATION OF SLUICE GATES AND ITS APPLICATION

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(Paper in Vol. 2 p. 1526)

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