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THE PERMAFROST REGIME
IN THE MACKENZIE DELTA-BEAUFORT SEA
REGION, N.W.T.
AND ITS PALEOCLIMATIC IMPLICATIONS

by

Diana M. Allen

A thesis submitted to the Faculty of
Science in partial fulfillment of
the requirements for the degree of
Master of Science

Department of Earth Sciences

Carleton University

Ottawa, Ontario

March, 1988

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The undersigned hereby recommend to the
Faculty of Science acceptance of this thesis, submitted
by Diana Allen, in partial fulfillment of the requirements
for the degree of Master of Science.

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ABSTRACT

Observed depths to the base of the ice-bearing permafrost (IBPF) as determined from geophysical well-log and survey information in the Mackenzie Delta and offshore Beaufort Sea continental shelf show significant variation. Depths range from over 700m beneath the northeastern portion of Richards Island and the adjacent offshore region to the north, to less than 100m in the modern Mackenzie Delta and offshore Mackenzie Bay regions. Along the Tuktoyaktuk Peninsula, permafrost thicknesses exceed 600m but tend to thin to less than 100m in a southeasterly direction.

The wide variation in the thickness and distribution of permafrost is due in part to differences in the subsurface deltaic lithologies and to the complexity of the past surface temperature history, exhibited, for example, by fluctuations of the mean annual ground temperature. Geological cross-sections showing variation in the Tertiary and Upper Cretaceous deltaic sequences indicate a good correlation with the permafrost distribution. In particular, where the sand and gravel units of the Iperk and Kugmallit Sequences predominate, permafrost is thick owing to the high thermal conductivities of the ice-rich sediments. Conversely, where the low thermal conductivity shales and mudstones of the Richards Sequence are present, permafrost is accordingly thinner.

Paleoclimatic reconstructions for the Mackenzie Delta using numerical models of permafrost aggradation indicate that permafrost has aggraded since the end of the Sangamonian (75,000 years B.P.)

when surface temperatures were approximately -1°C . An Early and Late Wisconsinan surface temperature of -18°C and a Mid-Wisconsinan rise to -8°C , similar to mean surface temperature today, are sufficient to account for observed maximum permafrost thicknesses in the region.

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This thesis is based in part on permafrost data made available by the former Earth Physics Branch and on results of observations reported in various papers on the Mackenzie Delta region.

The research was originally proposed by Dr. Alan Judge, Geological Survey of Canada whose assistance and guidance are gratefully acknowledged. Special thanks is extended to Dr. Frederick Michel, Department of Earth Sciences, Carleton University, for his enthusiastic support and supervision throughout the course of this study.

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The Mackenzie Delta region is characterized by ground temperatures which range from as low as -10°C on the Tuktoyaktuk Peninsula to just below 0°C in the vicinity of the modern Mackenzie Delta. Figure 2 (after Young and Judge, 1985) shows the variation of mean annual ground temperatures in the Mackenzie Delta region. This wide variation is due primarily to the presence or absence of bodies of water - a factor which has a strong influence on the ground surface temperature.

The surface topography and geomorphology of the Mackenzie Delta region, which is composed in part by modern sediments - forming the modern delta - and older deltaic and fluvial deposits - composing Richards Island, the offshore Beaufort shelf and the Tuktoyaktuk Peninsula - have been described in detail by Mackay (1963, 1974).

The Mackenzie Delta region covers an area which extends from about 130° to 137° longitude and from 68° to 71° latitude. It is centered around the modern Mackenzie Delta which is approximately 70 kilometres wide and 140 kilometres long. The modern delta is a low, flat area interlaced and dissected by several large channels and numerous smaller meandering channels, and is spotted with thousands of lakes and ponds. To the west it is bordered by the Richardson Mountains consisting predominantly of sedimentary formations of Mesozoic age; to the east by Richards Island and the Caribou Hills consisting of extensive formations of Paleozoic age covered in

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CHAPTER 1-INTRODUCTION

1.1 Aspects of Permafrost

The Mackenzie Delta, N.W.T., located in the western Canadian Arctic, forms the terminus of the Mackenzie River at the Beaufort Sea coast (Figure 1). Its northern location coupled with the geomorphic characteristics of the surface and the thermal regime at depth have provided suitable conditions for the development of permafrost which typically extends to depths of hundreds of metres.

The word Permafrost implies ground which remains frozen; although under changing natural or man-induced conditions it may thaw. Under these conditions the frozen state is not permanent, hence permafrost is best described as ground which remains frozen (below 0°C) throughout the summer, and thus through the succeeding winter. The depth at which the ground temperature is measured is necessarily below the near-surface layer where annual temperature fluctuations occur. This depth is referred to as the depth of zero annual amplitude.

In northern regions, the active layer - that layer which freezes and thaws on an annual basis - is generally thin as compared to its thickness in more temperate regions. However, this does not preclude the fact that annual thawing and refreezing of the near-surface frozen sediments is one process responsible for the moulding of the natural landscape; thermokarst and frost heave are two examples of such processes.

The mean annual ground temperature generally increases very

slowly with depth due to the loss of heat from the earth's interior. The rate of this increase is referred to as the geothermal gradient. Towards the ground surface, the mean annual ground temperature (MAGT) is determined by ground climate and the nature of the heat energy transfer processes through the ground surface. The MAGT is normally different, and is generally warmer than the mean annual air temperature (MAAT) by up to several degrees. Furthermore, this difference is by no means constant, but varies with the particular nature of the ground surface - that is, whether grass-covered or forest, mountainous or tundra - and the microclimate.

On a small scale, a detailed analysis of the microclimate of a region - which would include an analysis of those factors which contribute to the geomorphic conditions of the ground surface - is essential to a complete understanding of the permafrost regime. More specifically, towards the southern limits of permafrost, in the 'discontinuous' permafrost zone, the thickness and lateral extent of the permafrost varies greatly within short distances. This type of occurrence is a reflection of small variations in the microclimate, for example, relief, vegetation, slope exposure; hence a complete evaluation of the region would require a precise knowledge of the mean ground temperatures at various site locations. This would ultimately depend on the climatic and surface conditions affecting the energy exchanges between the ground and the atmosphere.

On a much larger scale, such as the entire Mackenzie Delta region, where permafrost is extensive and deep, a detailed investigation of the permafrost occurrence in terms of variations in the microclimate as described above would not be a practical approach. Deep ground temperatures are a reflection of an averaged

mean annual ground temperature (MAGT). Variations in ground surface conditions, such as the presence or absence of large bodies of water, subsurface lithologic conditions and long-term fluctuations in surface temperature are the major factors which have a bearing upon the distribution and thickness of deep permafrost.

The Mackenzie Delta region is characterized by ground temperatures which range from as low as -10°C on the Tuktoyaktuk Peninsula to just below 0°C in the vicinity of the modern Mackenzie Delta. Figure 2 (after Young and Judge, 1985) shows the variation of mean annual ground temperatures in the Mackenzie Delta region. This wide variation is due primarily to the presence or absence of bodies of water - a factor which has a strong influence on the ground surface temperature.

The surface topography and geomorphology of the Mackenzie Delta region, which is composed in part by modern sediments - forming the modern delta - and older deltaic and fluvial deposits - composing Richards Island, the offshore Beaufort shelf and the Tuktoyaktuk Peninsula - have been described in detail by Mackay (1963, 1974).

The Mackenzie Delta region covers an area which extends from about 130° to 137° longitude and from 68° to 71° latitude. It is centered around the modern Mackenzie Delta which is approximately 70 kilometres wide and 140 kilometres long. The modern delta is a low, flat area interlaced and dissected by several large channels and numerous smaller meandering channels, and is spotted with thousands of lakes and ponds. To the west it is bordered by the Richardson Mountains consisting predominantly of sedimentary formations of Mesozoic age; to the east by Richards Island and the Caribou Hills consisting of extensive formations of Paleozoic age covered in

places with glacial drift and moraine deposits; and to the northeast by older deltaic or fluvial deposits (Mackay, 1963).

1.2 The Permafrost Regime of the Mackenzie Delta

The Mackenzie Delta has in recent years been a center of northern scientific research and hydrocarbon exploration. Increased drilling activity in both onshore and offshore regions during the 1970's has provided the means to map the occurrence of deep permafrost. Conventional logging information and surveys acquired during exploratory drilling for oil and gas have been successfully used to establish the depth to the base of the ice-bearing permafrost (IBPF) and hence permafrost thickness. This has been made possible by combining a knowledge of the physical properties of permafrost and a recognition of the characteristic responses of conventional well-logging instruments to frozen ground (Osterkamp and Payne, 1981; Hatlelid and MacDonald, 1979; Hnatiuk and Randall, 1977). In addition, the measurement of subsurface temperatures in abandoned oil and gas wells - reported in the Canadian Geothermal Data Collections for Northern Wells (Taylor and Judge, 1977; Judge et al., 1979; Judge et al., 1981; Taylor et al., 1982) - has provided calibration of the log responses and a general idea of the thermal regime at depth in the Mackenzie Delta.

The Mackenzie Delta region is to a large extent underlain by thick continuous permafrost. Unfrozen sections occur in newly deposited unconsolidated sediments where the terrestrial climate has just begun to impose its influence on the ground thermal regime, and beneath major rivers and lakes. The unfrozen zones are generally

referred to as taliks and are the result of local surface warming derived from the existing bodies of water or localized climatic disturbances. Beneath these taliks, the permafrost usually resumes its continuity and attains a depth which reflects the combined effect of subsurface lithologic variations, past surface temperature fluctuations and the present surface conditions.

Permafrost is also absent in certain offshore regions which have remained submerged for sufficient lengths of time - such as shelf areas where water depths exceed 80 metres - or which have been acted upon by higher surface temperatures. The latter situation is best described by the Mackenzie Trough area where repeated glaciation during the Pleistocene (Blasco, personal communication) and substantial sediment accumulation during the Holocene have caused a warming in sub-seabottom temperatures.

Although permafrost is absent or relatively thin beneath certain disturbed environments in the Mackenzie Delta, it is in general very extensive and very deep - exceeding 700m in some localities. Onshore, downhole logging surveys have provided detailed information concerning the ice-bearing permafrost thickness and distribution over much of the region. These well-log studies carried out by the D&S Petroleum Consulting Group (1983) and summarized by Judge and Taylor (1985) are sufficient to describe large-scale trends in the occurrence of permafrost.

Offshore, seismic and temperature observations in the Beaufort Sea (Neave et al., 1978; Hunter et al., 1976; Weaver and Stewart, 1982) indicate extensive ice-bonded sediments beneath much of the Beaufort continental shelf. This offshore permafrost is best described as relic degrading permafrost, as its distribution is a

6.
reflection of the past sub-aerial landscape whose present thicknesses have been arrived at by gradual degradation due to the recent warming effect of submergence.

Taking into account the varied occurrence of permafrost throughout the region, it would seem that a number of different factors have contributed to the formation and preservation of the permafrost regime. It is the combined effect of these factors, which have been briefly discussed in the previous section, which will be discussed in detail in order to establish the formational history of this unique permafrost environment.

1.3 Objectives of Research

Previous research carried out by the author and reported in Leblanc-Ross (1986) gave a clear indication that subsurface variations in lithology coupled with a complex surface temperature history during the Late Pleistocene and Holocene were sufficient to account for the observed occurrence of permafrost in the Mackenzie Delta. These two parameters were examined qualitatively in order to establish their relative effect on the permafrost regime.

As expected, large-scale trends in the subsurface lithologies showed a strong correlation with the permafrost distribution. Where high thermal conductivity sandstones dominated the subsurface, permafrost was thick, and where low thermal conductivity shales and mudstones dominated, permafrost was accordingly thinner.

The thickness of permafrost was observed to be strongly dependent upon the past surface temperature history, in the sense that areas which remained emergent and unglaciated during the

greater part of the Late Pleistocene - such as the northeastern portion of Richards Island - tended to support thicker permafrost. Alternatively, in areas which were extensively glaciated or which were submerged beneath the Beaufort Sea for a sufficient length of time, much thinner permafrost was present. The modern Mackenzie Delta and the offshore Mackenzie Bay regions are two examples where permafrost is thin to absent as a result of repeated glaciations and lengthy periods of submergence.

These marked correlations have provided grounds for subsequent research of permafrost occurrence in the Mackenzie Delta region. The present objective is to measure quantitatively the effect of variations in lithology in terms of variations in thermal conductivity as well as the effect of fluctuations in surface temperature history in order to reconstruct a paleoclimate which, under the lithologic constraints, will account for the observed permafrost regime.

By examining the lithologic character of the subsurface and constructing a three-dimensional view of the variations in rock type throughout the Mackenzie Delta, sub-regions of similar lithology may be designated. Through the use of published temperature data for the subsurface, estimates of the thermal conductivities for the various sequences will be made and used in modelling different climatic histories of the sub-regions.

Using the knowledge of permafrost occurrence, the models for paleoclimate will be discussed in terms of their viability as estimates of the actual paleoclimate of the region.

In the following pages, details concerning the permafrost regime of the Mackenzie Delta, the regional geology, the thermal

aspects of the problem, the climatic factors which have in the past been and are at present active in the region, and the modelling techniques will be discussed. These topics will serve as important background information for the final evaluation of the models and discussion of a possible paleoclimatic history.

FIGURE 1 The Study Area.

CHAPTER 2 - PHYSICAL ASPECTS OF PERMAFROST

Ground that remains below 0°C for more than two consecutive years is referred to as Permafrost and as such, only accurate temperature measurements can define its boundaries. In contrast, frozen ground is a result of the freezing of aqueous solutions which occupy the pore spaces in rocks and soils, and, therefore, may be referred to therefore as Ice-bearing Permafrost (IBPF).

The temperature at which the ground becomes ice-bearing is lower than 0°C by a factor called the freezing point depression (FPD). The FPD varies in response to differences in pressure, - particularly on a microscopic scale - salt concentration and grain size. Consequently, water present in the ground at temperatures below 0°C may be only partially frozen and changes in the physical properties will occur over a range of temperatures below 0°C rather than precisely at 0°C . It is for this reason that the depth to the 0°C isotherm will not coincide with the IBPF base; in fact, these will only coincide for pure water at atmospheric pressure.

2.1 Permafrost Detection Techniques

The most reliable technique used in detecting permafrost is the temperature survey. Accurate temperature data acquired along the length of a borehole subsequent to drilling often provide a good estimate of permafrost thickness. Temperature profiles obtained shortly after the completion of a well may exhibit a distinct increase in temperature by several degrees at the base of the frozen section. This increase is due to the sudden drop in ice content

which becomes apparent as a result of the combined effects of latent heat in the ice and borehole heating during drilling. This step is interpreted to be the base of the IBPF, and is generally less than the depth to the 0°C isotherm for the reasons noted above.

The determination of the depth to the base of the ice-bearing permafrost is often not possible, either because the early logs are not available or ice contents are very low. The gradual transition from frozen to unfrozen observed in fine-grained materials reduces the prominence of the temperature step.

In cases where the early temperature logs are not available, the return to equilibrium of the well may be monitored over a number of years through the use of permanently installed temperature cables or portable logging cables. Temperature gradients in unfrozen materials of high porosity are greater than those in the frozen sections and a distinct increase in gradient is sometimes observed. This gradient change marks the base of the frozen layer. Where no such gradient change is apparent, temperature data serves as a precise indication of the depth to the zero degree isotherm.

IBPF manifests itself in the earth's measurable physical properties. Significant changes in the electrical and acoustical properties are observed as a result of freezing (Aptikaev, 1964; Pandit and King, 1979; Kurfurst and King, 1972; Timur, 1968; Hoekstra and McNeill, 1973) and hence, techniques used to measure such properties are commonly used in permafrost investigations (Osterkamp and Payne, 1981; Hatlelid and MacDonald, 1979; Hnatiuk and Randall, 1977).

As the temperature drops below 0°C in the materials in question, water confined in the pore spaces is transformed into ice

and the acoustic velocity undergoes a sharp, non-linear increase. This response is generally more apparent in the coarse-grained materials such as sands and gravels. Under these conditions, the acoustic velocity technique is useful in identifying the frozen zones.

In addition to temperature and lithology, acoustic velocity is also dependent upon variations in porosity, ice content and pore water salinity. This multi-dependence renders acoustic velocity techniques less reliable to accurately determine the depth to the base of the IBPF.

Two major wireline tools used to measure acoustic velocities are the Borehole Compensated Sonic and Crystal Cable surveys. The former log is a recording of the time required for a compressional wave to travel through a unit length of formation. Since transit time is the reciprocal of velocity, permafrost would be characterized by short travel times. Crystal cable surveys (discussed in detail by Walker and Stuart, 1976) involve the use of a series of detectors or geophones which are suspended in the well at pre-determined intervals. A velocity plot is obtained through the use of arrival times at the geophones from shot holes on the surface. Despite the usefulness of this technique in determining the IBPF base, it should be noted that when estimates of the IBPF base determined using this technique are compared to estimates obtained using other geophysical techniques, the former are invariably shallower. This may be an indication that crystal cable surveys underestimate the IBPF base.

In general, the response of electrical resistivity is much more pronounced than that of acoustic velocity. The former has been

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reported to undergo increases of up to 300 times during the transition from the unfrozen to the frozen state; whereas acoustic velocities may only increase 2 to 10 times during the same transition (Hnatiuk and Randall, 1977).

As the temperature drops below 0°C, the electrical resistivity undergoes a rapid increase - this being more marked in the coarser-grained materials. In fine-grained materials, such as clays, silts and shales, a gradual transition from unfrozen to frozen states is observed. This type of response is believed to be related to unfrozen water content which may be quite significant in the finer-grained materials. Williams (1982) discusses the effects of grain size and temperature on the unfrozen water content in frozen soils.

Geophysical techniques used to measure electrical properties of frozen ground include induction, spontaneous potential (SP) and electromagnetic probing. Induction logs record the resistivity of the subsurface materials. The Dual Induction-Log (DIL) has proven to be the most useful of the wireline tools for detecting the presence of frozen ground. In fact, it is the most widely relied upon technique for determining the depth to the base of the IBPF in the Mackenzie Delta area.

SP records the natural potential difference between a moveable electrode in the borehole and a fixed electrode at the ground surface. In the permafrost zone, the SP curve shows a characteristic negative deflection - the start of this negative shift coinciding with the base of the permafrost in a number of wells.

Electromagnetic probing was not used to establish the permafrost base in this particular database, however, it is a

noteworthy technique because surface soundings have been used and is discussed in detail by Hoekstra and McNeill (1973).

In addition to the above mentioned techniques, there are various other geophysical methods which have been successfully used to confirm or estimate the depth to the base of the IBPF. Reflection seismic surveys have been widely used to delineate permafrost zones (Hnatiuk and Randall, 1977) - particularly in the offshore regions (Hunter et al., 1976). As well, caliper surveys run to detect overgauge holes caused by caving may be used indirectly since caving does not occur in competent permafrost formations (Osterkamp and Payne, 1981).

Other useful logs which including neutron and density porosity have also been used. These logs, however, are often difficult to interpret in isolation.

2.2 The Database

The interpretation of a large number of logs of the types discussed above for 170 wells in the Mackenzie Delta has been carried out and summarized by Judge and Taylor (1985), and forms the basis of the database used for analysis here. Appendix I includes a listing of the site locations along with estimates of the depth to the 0°C isotherm where data were available and the depths to the base of the IBPF.

Columns 1 through 3 of this table are self explanatory. Column 4 lists the available estimates of the depth to the 0°C isotherm and the method by which it was determined. These data are reported in the Canadian Geothermal Data Collections for Northern Wells (Taylor

and Judge, 1977; Judge et al., 1979, 1981; Taylor et al., 1982).

Although few in number, these data constrain the interpretation of the geophysical log responses, and in some cases - particularly where "less than" values for the base of the IBPF are quoted - provide the only estimate of the permafrost thickness.

The symbol preceeding the 0°C isotherm depth indicates the technique used to establish this depth; "E" indicates Equilibrium thickness which means that the value in question was determined from the equilibrium temperature data. Equilibrium data probably provide the most accurate estimates of the permafrost thickness.

An "X" indicates an extrapolated thickness. In these cases, temperature data were not available down to the depth in question and the temperature profile was extrapolated to obtain the 0°C isotherm depth. Variations in lithology below the available temperature data may alter gradients, hence extrapolated values - which do not take these variations into account - are less reliable.

A "greater than" (>) symbol indicates that data were unsuitable for interpretation and a minimum value is provided. If the depth to the 0°C isotherm is followed by a "+" symbol, this indicates that the value is probably underestimated.

Column 5 provides the best estimate of the depth to the base of the ice-bearing permafrost. Those values preceeded by a "less than" (<) symbol indicate that the depth to the base of the IBPF occurs above the depth where the geophysical log begins. A number of these values have been modified from the original database (Judge and Taylor, 1985) to incorporate selected geophysical data reported in the Canadian Geothermal Data Collections for Northern Wells. As well, IBPF depths determined by the crystal cable technique are

suspected to fall short of the actual values. Where possible these were replaced by depth estimates which are similar to those of the surrounding areas.

The final column indicates the geophysical technique used to estimate the depth to the base of the IBPF. These include Resistivity (Res), Crystal Cable (Cry), Acoustic Velocity (Ac) and Open Hole Temperature (OHT).

A point worth noting is that the geophysical logs used to estimate the depth to the base of the IBPF were obtained shortly after drilling and were run from the well-drilling platform (Kelly Bushing). Temperature logs obtained after the wells have been abandoned are taken from ground level. Therefore, there exists a slight discrepancy between the two estimates of the permafrost thickness. The average difference between the Kelly Bushing elevation and ground surface elevation is $7.3\text{m} \pm 2.1\text{m}$ and this value should be subtracted from the IBPF bases when they are being compared to the 0°C isotherm depths.

With this in mind, it is possible to draw some conclusions respecting the correlation between the depth to the base of the IBPF and the 0°C isotherm depth. As mentioned earlier, the 0°C isotherm would be expected to be deeper than the depth to the base of the frozen section because of the role played by the freezing point depression. Upon examination of the table in Appendix I, this is clearly not always the case. Not infrequently is the depth to the 0°C isotherm shallower than the depth to the base of the IBPF. However, in most of these instances, the difference is small enough that if the Kelly Bushing elevation is subtracted from the latter, the two values are approximately the same.

Occasionally, there is a much greater difference between the two values which cannot be explained by the elevation difference alone. Judge (personal communication) suggests that gas hydrates at depth may be the cause of the discrepancy. Gas hydrates contained in or below the water-ice layer may confuse the base of the ice-bearing permafrost since they exhibit a similar log response, causing it to appear deeper than it actually is.

Despite these minor discrepancies, there appears to be no discrete pattern in the difference between the depth to the 0°C isotherm and the depth to the base of the IBPF, with the exception that the depth to the 0°C isotherm is generally greater. This fact should be taken into account when the results of thermal modelling are discussed. Because the models are based on temperature alone, the calculated depth to the base of the permafrost (0°C isotherm) will be consistently deeper than the database values which reflect the depth to the base of the frozen layer.

2.3 The Distribution of Onshore Permafrost

Although minor modifications were made to the database, the permafrost map provided in Judge (1986) is used to illustrate the variation in the thickness and distribution of ice-bearing permafrost in the Mackenzie Delta region (Figure 3). This map does not incorporate the available estimates of the depth to the 0°C isotherm nor does it take into account the few altered values of the depth to the base of the IBPF determined using the crystal cable survey. Despite these inconsistencies, the permafrost map provided here best illustrates the available data.

With reference to Figure 3, the thickest permafrost is observed beneath the northeastern portion of Richards Island where measured thicknesses exceed 700m. Further to the northeast, beneath Kugmallit Bay, permafrost thins very gradually but remains thick beneath much of the Beaufort Sea continental shelf. Beneath the western portion of Richards Island, there is a rapid decrease in the thickness of permafrost and beneath the modern Mackenzie Delta, permafrost is thin to absent.

Westward and along the Yukon Coastal Plain, permafrost may again thicken, however, the sparsity of data points in this region is insufficient to confirm this. Lachenbruch et al. (1982) have reported thicknesses of up to 600m at Prudhoe Bay, Alaska.

In the southern portion of the Mackenzie Delta, specifically in the Parsons Lake area, permafrost rapidly thins from over 600m to less than 200m in a southerly direction.

2.4 Distribution of Offshore Permafrost

Offshore, shallow seismic refraction and temperature observations in the southern Beaufort Sea (Neave et al., 1978; Hunter et al., 1976; Weaver and Stewart, 1982) indicate extensive ice-bonded sediments beneath much of the continental shelf. The velocity model applicable to the Beaufort Sea (Hunter et al., 1976) consists of a low velocity water layer (1420 ms^{-1}) overlying unfrozen bottom sediments ($1420\text{-}2000 \text{ ms}^{-1}$) which in turn overlies ice-bonded materials ($2000\text{-}4500 \text{ ms}^{-1}$). In general, the velocity of the frozen ice-bonded material is greater than that of the unfrozen state. Changes in magnitude of up to five times the unfrozen

velocity can occur. Figure 4a (after Neave et al., 1978) shows the distribution of high seismic velocities in the Beaufort Sea interpreted to be associated with ice-bonded sediments - velocities greater than 2500 ms^{-1} are interpreted as ice-bonded sediments.

The distribution of offshore permafrost is best illustrated in Figure 4b (after Hunter et al., 1976). In general, the offshore region may be subdivided into three zones: 1) continuous ice-bonding, 2) discontinuous ice-bonding and 3) no ice-bonding. These correspond to thick high seismic velocity layers, thin high seismic velocity layers and low seismic velocity layers, respectively.

No evidence of ice-bonding has been found offshore where water depths exceed 80 metres, thus the outer limit of offshore permafrost in the Beaufort Sea corresponds approximately to the 80m water-depth contour. In addition, ice-bonded sediments are absent from most parts of Mackenzie Bay and the Mackenzie Trough (Blasco, 1987).

The top of the ice-bonded permafrost is also variable, this being a reflection of past surface conditions. In the nearshore region (specifically to the northeast of Richards Island), the top of the permafrost is quite rugged with slopes of 10° or more. Further north, about 40km offshore, the top of the permafrost is at a depth of 150m below seabottom. Hunter et al. (1976) suggest that this drop may be associated with the interpreted "ancient" Mackenzie River channel outlined by seabottom topography.

Seismic profiles have proven invaluable in the delineation of offshore permafrost and there is a good correlation between the depths to the top of the ice-bonded permafrost as determined from available temperature and lithologic borehole logs, and from seismic

records.

In situations where no data are available on sediment temperature, grain size or pore water salinity, seismic compressional wave velocities may serve only as a qualitative estimate of ice content. The velocities of ice-bonded sands and gravels having no observed excess ice are, in general, much higher than those for massive ice lenses (Hunter, 1973).

Shearer et al. (1971) describe - by seismic reflection techniques - numerous underwater mounds mapped on the seabottom of the continental shelf. These dome-like structures are interpreted as pingos formed in ancient lake bottoms. Figure 5 (after Hunter et al., 1976) shows a typical pingo-like feature in the southern Beaufort Sea. With few exceptions, these sub-marine pingos are associated with the discontinuous ice-bonded permafrost zone.

Where present, offshore permafrost is best described as relic degrading permafrost as it exists in a state of disequilibrium with the seabottom. In fact, very little offshore permafrost is in equilibrium with negative seabottom temperatures. Its distribution is a reflection of the past sub-aerial landscape and its present thicknesses have been arrived at primarily by gradual degradation due to the recent warming effect of submergence. Despite the overall tendency towards degradation of permafrost in the offshore, there is evidence of aggradation at shallow depths in the sea floor - possibly marked for example, by the growth of features analogous to terrestrial pingos (Shearer et al., 1971).

2.5 Factors Controlling the Permafrost Regime

The variable occurrence of permafrost in the Mackenzie Delta region may be explained in a qualitative way, primarily by differences in the thermal properties of the subsurface lithologies and by the complex long-term surface temperature history, and to a lesser degree by variations in the terrestrial heat flux.

The thermal properties which control the formation and the existence of permafrost include thermal conductivity, heat capacity and thermal diffusivity. Variations in thermal conductivity have the most significant effect on the thermal regime, particularly under steady-state conditions. The relationship between thermal gradient, thermal conductivity and heat flow in one dimension is defined by the equation:

$$Q = K dT/dZ, \quad (1)$$

where Q is the steady-state heat flow, K is the thermal conductivity and dT/dZ is the thermal gradient.

While heat flows as a result of a temperature difference, and the amount increases in proportion to the difference, the rate of heat transfer depends on the thermal conductivity of the material through which the heat flows. Materials - in this case rocks and sediments in the subsurface - which have high thermal conductivities will display smaller temperature gradients and, therefore, will support thicker permafrost.

The relationship between thermal conductivity - as exhibited by variations in the subsurface lithologies - and the occurrence of permafrost will be discussed at a later point.

Although not as important as thermal conductivity to the

maintenance of permafrost, heat capacity and thermal diffusivity are significant properties to consider in its formation.

Heat capacity determines the amount by which the temperature of a body will change in response to heat being transferred through it. The higher the heat capacity of a rock, the greater the amount of heat required to raise the temperature of that rock. It should be noted that the heat capacity of the water content forms the largest component of the total heat capacity; therefore, an estimate of porosity is essential.

The ratio of the thermal conductivity to the heat capacity defines the thermal diffusivity, α . This property is essential when modelling the transient response of the subsurface to a temperature disturbance at the surface.

A more detailed discussion of the thermal properties and their effect on the permafrost occurrence is given in Chapter 4.

Surface temperature history is the second major factor which affects the permafrost regime. Major contributions to the surface temperature history include fluctuations in mean annual air temperature - manifested as changes in the mean annual ground surface temperature, the extent and duration of glacial cover, fluctuations in sea level associated with the submergence and emergence history of the region, and shoreline retreat which has been active along the shorelines during the past few thousand years. Details concerning the surface temperature history of the Mackenzie Delta region will be discussed in Chapter 5 and will serve to define the parameters used for thermal modelling.

The final condition which is of importance in determining the occurrence of permafrost is the geothermal heat flux, Q . Variations

in heat flow throughout the region will have a significant effect on the distribution and thickness of permafrost. In the present analysis, it is assumed that the terrestrial heat flow remains constant and uniform throughout the region, and, therefore, the permafrost regime is controlled primarily by only surface temperature history and ground thermal properties.

D

FIGURE 3 Depth to the Base of the Ice-Bearing Permafrost in the Mackenzie Delta as Determined from Industry Well Logs and Government Temperature Surveys (after Judge, 1986).

sediments are divisible into two main types: mud-dominant and sand-dominant, corresponding to pro-delta and delta-plain depositional environments, respectively. As such, lithologic differentiation may be applied. However, basinward the succession is lithologically similar over a considerable thickness of strata, and sequence nomenclature is more suitable.

Dixon (1986) identified eleven sequences using reflection seismic profiles, well logs and, to a limited extent outcrop data. From the youngest to the oldest these are: Shallow Bay, Iperk, Akpak, Mackenzie Bay, Kugmallit, Kopanoar, Richards, Reindeer, Fish River, Smoking Hills and Boundary Creek. A description of the lithologies comprising each sequence (after Dietrich et al., 1985) is described below and is briefly summarized in Table 1.

3.1.1 Shallow Bay Sequence

The Shallow Bay Sequence is the youngest identifiable seismic sequence. It is best developed in the Mackenzie Trough where it reaches an estimated thickness of 400m. Outside the trough, the sequence is thinner and not detectable.

The Shallow Bay Sequence includes sediments of the modern Mackenzie Delta and is thought to consist of mud, silt, sand and gravel (Vilks et al., 1979). It is assumed that the deposition of this sequence began after the last major glaciation and is mostly Holocene, however, sediments as old as Late Pleistocene may be included.

3.1.2 Iperk Sequence

The Iperk Sequence forms a distinct, thick lens-shaped deposit in which there is very little seismically visible structure. Its thickness varies from a few tens of metres at the basin margin to

FIGURE 4b An Interpretation of the Occurrence of Sub-Seabottom Ice-Bonded Permafrost from Industry Seismic Records (after Hunter et al., 1976).

FIGURE 5 Acoustic Reflection Profiles over Pingo-Like Features in the Southern Beaufort Sea (after Hunter et al., 1976).

CHAPTER 3 - REGIONAL GEOLOGY

Geological studies have been conducted in the Mackenzie Delta region since the early days of exploration; however, most of the regional geologic work began during two Geological Survey of Canada reconnaissance operations - "Operation Norman" and "Operation Porcupine". In the Mackenzie Delta, the work of Young et al. (1976), Halkings and Hatlelid (1975), Young and McNeil (1984), Dixon (1982) and Dietrich et al. (1985), provide the most comprehensive stratigraphic studies. As a result of these works, a number of informal and formal stratigraphic nomenclature schemes for the Upper Cretaceous to Holocene succession exist. Some successions have several names, or similar terms are used for different stratigraphic units.

Leblanc-Ross (1986) incorporated the lithostratigraphic nomenclature (identified as formations) proposed by Young and McNeil (1984) for a permafrost investigation in the Mackenzie Delta. Since that time, Dixon (1986) has suggested that the recognition of "sequences" is the most practical method for dividing the succession in the Beaufort-Mackenzie Basin. In accordance with Dixon (1986), a sequence nomenclature will be referred to in the present permafrost investigation.

3.1 Geological Framework

Late Cretaceous to Recent sedimentation was and still is dominated by deltaic processes. This has resulted in a series of thick, northward prograding delta-complexes. In general, the deltaic

sediments are divisible into two main types: mud-dominant and sand-dominant, corresponding to pro-delta and delta-plain depositional environments, respectively. As such, lithologic differentiation may be applied. However, basinward the succession is lithologically similar over a considerable thickness of strata, and sequence nomenclature is more suitable.

Dixon (1986) identified eleven sequences using reflection seismic profiles, well logs and, to a limited extent outcrop data. From the youngest to the oldest these are: Shallow Bay, Iperk, Akpak, Mackenzie Bay, Kugmallit, Kopanoar, Richards, Reindeer, Fish River, Smoking Hills and Boundary Creek. A description of the lithologies comprising each sequence (after Dietrich et al., 1985) is described below and is briefly summarized in Table 1.

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3.1.2 Iperk Sequence

The Iperk Sequence forms a distinct, thick lens-shaped deposit in which there is very little seismically visible structure. Its thickness varies from a few tens of metres at the basin margin to

about 5000m at its center of deposition in the outermost region of the shelf. Rapid thickening of this sequence begins just north of Richards Island and can be traced eastwards through Kugmallit Bay. In the western Beaufort Sea, the Iperk Sequence is relatively thin. The basin margin sediments consist of sand and gravel and are predominantly non-marine to marginal marine deposits near the Tuktoyaktuk Peninsula, grading laterally into marine sediments. The age of the Iperk Sequence is Pliocene-Pleistocene, although it may be as old as Late Miocene.

3.1.3 Akpak Sequence

The Akpak Sequence is present throughout most of the Mackenzie Basin. It is 700 to 800m thick in the area just north of Richards Island, but is thickest in the western Beaufort Sea where it is estimated to be 2500m thick. Its southern limit is a truncation edge which is located just north of Netserk B-44 and Pullen E-17 (refer to Appendix I).

Where penetrated, the Akpak Sequence is predominantly mud and silt with a few thin sand beds. The age of this sequence is probably Middle to early Late Miocene.

3.1.4 Mackenzie Bay Sequence

The Mackenzie Bay Sequence is present under most of the Beaufort Shelf. It is about 360m thick at Netserk B-44, thickening northward to about 1000m. Its thickest development is in the western Beaufort where it is estimated to be as thick as 2500m.

Where penetrated, the Mackenzie Bay Sequence is dominated by mud and silt, although some thin sand units are present in the Netserk-Tarsuit area. The age of this sequence is Late Oligocene to Middle Miocene.

3.1.5 Kugmallit Sequence

The Kugmallit Sequence underlies conformably, but abruptly, the Mackenzie Bay Sequence. It is present under much of the Beaufort Shelf and part of the Mackenzie Delta. It is up to 3000m thick at its center of deposition in the area to the immediate north of Richards Island. Northward and westward, there is depositional thinning; however, its southern and southeastern limits are erosional. The sequence is present beneath the northern edge of the delta and is present beneath the Iperk Sequence between the Taglu and Kumak areas.

In its center of deposition, the Kugmallit Sequence consists primarily of interbedded sand and mud of delta-front origin. To the south, the succession becomes sandier as delta-plain deposits become more prevalent. East and west of the delta-plain deposits, shelf and pro-delta muds are dominant. To the north of the deltaic and shelf deposits, deepwater muds and silts deposited as slope and submarine fan deposits are present. The age of the Kugmallit deposits are believed to be as young as Early Oligocene (Young and McNeil, 1984; Dixon et al., 1984).

It should be noted that much of the Beaufort Formation described by Young and McNeil (1984) cannot be differentiated from the uppermost strata of the Kugmallit Sequence. Consequently, most of the sands and gravels identified as Beaufort deposits by Young and McNeil (1984) are now regarded as Kugmallit beds.

3.1.6 Kopanoar Sequence

As the sediments assigned to the Kopanoar Sequence are found at depths exceeding about 3000m and hence are of little consequence to permafrost studies, only a brief description of their lithology and

age will be given. The Kopanoar Sequence consists mainly of silt and mud, but scattered throughout are sand-rich zones. The age of the sequence is estimated as Late Eocene or perhaps Early Oligocene.

3.1.7 Richards Sequence

The Richards Sequence is estimated to be 1200-1500m thick in the Ivik area, and up to 2000m thick in the Ellice area. Under the Mackenzie Delta, the sequence is present beneath the Iperk Sequence; in the offshore region, it is present beneath the Kugmallit Sequence and appears to be absent in the farthest offshore wells.

The dominant sediments are mud and silt, although local sand and gravel beds are present. The age of this sequence is Middle to Late Eocene (Young and McNeil, 1984).

3.1.8 Reindeer Sequence

Beneath Richards Island and the near offshore region, the Reindeer-Richards contact is conformable but abrupt. In the Caribou Hills along the southern border of the Mackenzie Delta and along most of the Tuktoyaktuk Peninsula, the Reindeer Sequence rests unconformably on the lower part of the Fish River Sequence and, in places, on Smoking Hills strata. In the western Beaufort, the Reindeer rests unconformably on the Fish River Sequence.

The main center of deposition of the Reindeer Sequence appears to have been located in the Ellice area, where more than 4000m of sediment may be present. Non-marine and marginal marine sands and gravels are present along the Tuktoyaktuk Peninsula. North and northwest of the delta center of deposition, the sequence is represented by pro-delta/shelf muds and silts. Northeast of the Taglu area, the sequence has not yet been penetrated; however based on the known facies distribution, it is believed that these areas

contain pro-delta muds and silts which grade northward into deepwater deposits. The age of the Reindeer Sequence is Paleocene to Early Eocene (Young and McNeil, 1984).

3.1.9 Fish River Sequence

Formational units contained in this sequence include the Moose Channel (exclusive of the Ministicog Member), Tent Island and Mason River. In the northern Richardson Mountains and southwestern Mackenzie Delta, the sequence is 1800-2000m thick and in the Natsek area it is 4000-6000m thick. The main center of deposition is located in the Beaufort-Mackenzie Bay area. The sequence is characterized by pro-delta mud and silt in its lower part - namely the Tent Island and Mason River Formations - and is overlain by a thick succession of delta-plain and delta-front sediments. The age of this sequence is Late Campanian to Paleocene (Young et al., 1976).

3.1.10 Boundary Creek and Smoking Hills Sequences

Both the Boundary Creek and Smoking Hills Sequences are relatively thin - only a few hundred metres or less - and can only be identified on basin margins.

The Boundary Creek Sequence consists of organic-rich shale with bentonites and is present in a few wells on the northernmost edge of the Richardson Mountains - for example, Fish River B-60. On the east side of the Mackenzie Delta, beneath the Tuktoyaktuk Peninsula and the Anderson Plains, the Smoking Hills Sequence rests unconformably on a variety of Paleozoic and Lower Cretaceous units. It consists primarily of bituminous shales.

The Boundary Creek and Smoking Hills Sequences represent the basal beds of the Upper Cretaceous to Holocene strata; however,

their extent throughout the basin is poorly known.

3.2 Lithologic Variations

Lithologic variations and permafrost thickness in the Mackenzie Delta may be illustrated with reference to a series of seven geological cross-sections. These are located on the map shown in Figure 6a and appear in Figures 7a through 7g. The geological information used to construct these cross-sections - which were drawn by linking borehole data by straight lines - was compiled from various geological reports on the region (Jones et al., 1980; Nentwich and Yole, 1982; Young and McNeil, 1984; Dietrich et al., 1985; Dixon et al., 1985; Dixon, 1986) and from borehole lithologies supplied by the Institute of Sedimentary and Petroleum Geology, Calgary, Canada.

In the legend shown in Figure 6b - which accompanies the cross-sections - four lithofacies have been identified. In general, these facies directly correspond to a particular sequence. However, sequences which are of a lesser importance to the present study - that is, those which are at sufficient depth to escape any association with permafrost - are grouped under one particular facies.

Variation in lithofacies appear to show a correlation with the observed regional variation in permafrost distribution. In particular, as shown in the cross-sections, there exists a fairly good correlation between the upper sand-gravel unit of the Pliocene-Pleistocene Iperk Sequence and the occurrence of deep permafrost. In areas where this sequence is thick, the depth to the

base of the permafrost is significantly greater than in areas where this sequence is thin to absent. In addition, where the Iperk is underlain by the sand-dominant Kugmallit Sequence (cross-sections B-B' and E-E'), permafrost attains its greatest depths - in excess of 700m in some areas.

Conversely, permafrost is much thinner in areas where the Iperk Sequence is underlain by the mud-dominant Richards Sequence - for example, beneath the western portion of Richards Island (cross-sections C-C' and D-D'). In areas which are underlain by relatively thick Iperk and Reindeer sediments, the base of the permafrost is much more variable. For instance, the base of the permafrost gradually deepens towards the east from 200 to 600m (cross-section A-A'). Although the Iperk does not remain at a constant thickness, it does not show any trend towards the east. One possible reason for the steady increase in permafrost thickness in this direction lies in the very nature of a sequence. The main center of deposition of the Reindeer Sequence is located in the western portion of the delta, beneath the modern Mackenzie Delta. Here, deposits are composed primarily of marine sands, silts and muds. Towards the east, along the Tuktoyaktuk Peninsula, the sequence grades to marginal marine and non-marine sands and gravels.

At the northeastern tip of the Tuktoyaktuk Peninsula, permafrost again thins despite the presence of the sand and gravel deposits of the Reindeer Sequence. In this case, the Iperk Sequence thins abruptly where Upper Cretaceous rocks dip steeply toward the basin (cross-sections D-D', E-F and D-C'-G).

In the modern Mackenzie Delta area - which includes the area extending from south of Shallow Bay out into Mackenzie Bay -

permafrost is thin to absent. Thin Iperk sediments underlain by Reindeer sediments of primarily marine origin are one possible explanation for the thin occurrence of permafrost in this area. A more likely explanation is the surface warming caused by the Mackenzie River outflow.

Likewise, in the offshore outer-shelf region, recent submergence has led to gradual degradation of permafrost. This degradation of permafrost is clearly shown in cross-section D-C'-G. A more detailed discussion of the effects of submergence on the base of the permafrost will be given in Chapters 5 and 7.

3.3 Discussion

The strong correlation between the Iperk Sequence thickness and the permafrost distribution is very evident on the isopach map of the Iperk Sequence shown in Figure 8. An important point to note is that all units contained within the Iperk were included in the isopachs. Thus, the pro-delta/shelf mudstones and deepwater sediments - which do not appear to correlate with permafrost distribution - are included along with the upper sand and gravel unit of the Iperk. Cross-section D-C'-G shows the general trend in these units which indicates a general thinning of the upper unit towards the far offshore.

A second map was drawn to display some large-scale trends in the regional geology. Figure 9 shows the variation in the percentage of sand-dominant lithologies in the Mackenzie Delta region. Those sequences which are incorporated in the map are: 1) the upper sand-gravel unit of the Iperk Sequence, 2) the Kugmallit Sequence

and 3) the Reindeer Sequence.

Although this map does not show as clearly the relationship between the base of the permafrost and lithologic variation, it delineates the low sand content zone in the western portion of Richards Island. This zone - composed primarily of Richards Sequence muds - is partially responsible for the considerable thinning of permafrost in that area.

Elsewhere in the region, there is a high percentage of sand in the subsurface. The exceptions are the offshore region and the southern portion of the Tuktoyaktuk Peninsula where gradual thinning and rapid thinning of these units occur respectively. Again, a correlation with the occurrence of permafrost is noted.

A number of conclusions concerning the relationships between the lithologic variations and the permafrost distribution are suggested: 1) In general, there appears to be a strong correlation between the isopachs of the Iperk Sequence (Figure 8) and the depth to the base of the ice-bearing permafrost (Figure 3); 2) Although the Iperk Sequence is present in the western portion of the delta, namely the modern Mackenzie Delta, the warming effect due to the Mackenzie River outflow is probably responsible for the occurrence of thin permafrost in the area; 3) The thick Iperk deposits in the northeastern portion of Richards Island and the thick accumulation of Kugmallit Bay sands and gravels at depth combine to provide favorable conditions for permafrost aggradation; 4) Towards the east, along the Tuktoyaktuk Peninsula, the gradual thickening of permafrost may be accounted for by the general trend from the center of sediment deposition to marginal marine and non-marine Reindeer deposits; 5) Abrupt thinning of permafrost beneath the northeastern

tip of the Tuktoyaktuk Peninsula is probably related to the abrupt thinning of the Iperk Sequence at the basin edge; 6) A thick accumulation of Richards Sequence strata beneath the western portion of Richards Island may account for permafrost thinning in that area; 7) Degradation of offshore permafrost may be explained by the combined effect of the thawing of permafrost as a result of submergence and a trend towards marine and deepwater facies Iperk sediments in the offshore.

The observed correlation between lithology and the depth to the base of the permafrost is best described in terms of variations in the thermal properties of the subsurface rocks and sediments corresponding to lithological differences. A more detailed discussion of the thermal properties is given in the following chapter.

SYSTEM	SERIES	SEQUENCE	LITHOLOGY
QUATERNARY	HOLOCENE	SHALLOW BAY	Absent in Richards Island area, mud, silt, sand and gravel sediments.
			EROSIONAL UNCONFORMITY
	PLEISTOCENE	GLACIAL DEPOSITS	Silt, thin bedded, interbedded sand, pebbly silt, peat
			DISCONFORMITY
TERTIARY	NEOGENE	PLIOCENE	Basin margin sand and gravel, non-marine to marginal marine deposits Lateral gradation into marine sediments. Interbedded mud, silt, sand
			EROSIONAL UNCONFORMITY
		MIocene	AKPAK
			Mud and silt, with few thin sand beds.
	PALEOGENE		MACKENZIE
			Mudstone, light gray to brownish gray, slightly sandy, scattered chert pebbles.
		OLIGOCENE	KUGMALLIT
			Sand and gravel delta plain deposits, fine to coarse-grained chert and quartz. Delta front, interbedded sand and mud, shell and prodelta muds at depth to east and west of delta plain deposits.
			KOPANOAR
			EROSIONAL UNCONFORMITY
			Silt and mud, scattered sand-rich zones
		EOCENE	RICHARDS
			Mudstone, light gray, variably silty, sandy or carbonaceous
			REINDEER
			Non-marine and marginal marine sands and gravels. Prodelt / shell muds and silts to north and northwest of Tuktoyaktuk Peninsula. Ministieoog Member interbedded mudstone and siltstone
		PALEOCENE	FISH RIVER
			Moose Channel alternating sandstone and mudstone, delta plain front. Tent Island prodelta mudstone and siltstone Mason River shale.
UPPER CRETACEOUS	MAASTRICHTIAN		
	CAMPANIAN	SMOKING HILLS	DISCONFORMITY
	SANTONIAN		UNCONFORMITY
	CONIACIAN		Bituminous shale, probably highly radioactive.
	TURONIAN		UNCONFORMITY
	CENOMANIAN	BOUNDARY CREEK	Organic-rich shale, bentonites and concretary layers

TABLE 1 Table of Geological Sequences.

FIGURE 6a Map Showing Location of Lines of Cross-Section.

LEGEND

lithology

SAND AND GRAVEL, NON-MARINE
DELTAIC DEPOSITS - IPERK SEQUENCE

NON-MARINE AND MARGINAL MARINE
SAND AND GRAVEL - REINDEER SEQUENCE

SAND AND GRAVEL DELTA-PLAIN
DEPOSITS - KUGMALLIT SEQUENCE

MUDSTONES AND SHALES - RICHARDS,
MACKENZIE BAY, AKPAK, ARCTIC RED SEQS.

contacts

SEQUENCE BOUNDARY

FACIES CHANGE

UNCERTAIN SEQUENCE BOUNDARY

BASE OF ICE-BEARING PERMAFROST

FIGURE 6b Legend to Accompany Cross-Sections

FIGURE 7a 'Cross-Section A-A'

FIGURE 7a Cross-Section A-A' (cont'd)

FIGURE 7b Cross-Section B-B'

C'

C

FIGURE 7c Cross-Section C-C'

D

FIGURE 7d Cross-Section D-D'

FIGURE 7d Cross-Section D-D' (cont'd)

FIGURE 7e Cross-Section E-E'

FIGURE 7f Cross-Section E-F

FIGURE 7g Cross-Section D-C'-G

FIGURE 8 Isopach map of Iperk Sequence.

FIGURE 9 Map Showing the Variation in the Sand-Dominant Lithologies.

CHAPTER 4 - THE THERMAL REGIME

The thermal regime of the Mackenzie Delta is highly complex and variable in its nature. A modern mean annual ground temperature of -9°C (after Young and Judge, 1985) places the region in the zone of continuous permafrost. This, in itself, is a major contribution to the uniqueness of the thermal regime. In addition, the coastal location and lowland topography are responsible for the continued modification of the landscape and hence, the subsurface thermal regime. Lachenbruch et al. (1982) state that in coastal lowlands at high latitudes such as the Prudhoe Bay area - which is similar in many respects to the Mackenzie Delta - the thermal regime differs from that in more familiar temperate latitudes in two important respects: 1) Groundwater is generally immobilized in cold permafrost - which occurs primarily as ice - and thus, heat conduction theory can be applied with confidence to within a few metres of the surface, and 2) because the mean annual ground temperature is -9°C , bodies of water, such as lakes and rivers that do not freeze to the bottom (those greater than approximately 2m depth), constitute hot spots on the surface that contribute to perturbations of the subsurface temperatures.

With respect to the latter condition, the thermal interaction between bodies of water and the ice-rich permafrost on their shores causes lakes and rivers to coalesce, migrate and leave complex three-dimensional transient thermal disturbances in their wake (Lachenbruch et al., 1982; Taylor and Judge, 1983). As a result, the proximity of large bodies of water to permafrost sampling locations will have some influence on the observed permafrost depth. Lateral flow of heat must be considered and a two- or three-dimensional

model is necessary (Gold and Lachenbruch, 1973). As this information, concerning the proximity of large bodies of water, is known for only a small percentage of the datapoints, the significance of this effect has only been briefly noted.

4.1 Conductive Heat Transfer

In general, heat may be transferred by conduction, radiation and by mass transfer processes. Radiative transfer - in which heat is transferred directly between distant portions of a body by electromagnetic radiation - in rocks is only significant at temperatures higher than 1000°C (Judge, 1973) and, therefore, will not be discussed. Transfer of mass - which includes convective heat transfer - in the form of vapour or liquid in the presence of temperature gradients necessarily involves heat transfer. In partially saturated soils, this mode of heat transfer becomes important within a few metres of the surface. However, at depths of hundreds of metres, only groundwater circulation may cause modification of permafrost thicknesses and temperatures in limited areas.

The question of the effect of groundwater circulation in permafrost is debatable, particularly in regions of discontinuous permafrost or in areas which support numerous taliks. Water circulation in the unfrozen zones may be responsible for the slow degradation of permafrost in the surrounding zones as heat is transferred towards the latter (Smith and Hwang, 1973). This encroachment of taliks due to groundwater circulation may be very significant in the areas surrounding the Mackenzie River.

In continuous permafrost - where unfrozen zones are for the

most part absent - water may indeed migrate on a microscopic scale through ice-rich permafrost and massive ground ice, however, on a large scale - such as with regional permafrost variations - this mode of heat transfer is probably negligible.

The theory of heat conduction has been discussed in detail by Ingersoll et al. (1954) and by Carslaw and Jaeger (1959). A brief summary of some of the important aspects relevant to the present study are presented.

When different parts of a solid body are at different temperatures, heat flows from the hotter to the colder portions by the process of heat conduction. The rate at which heat will be transferred has been found by experiment to depend upon a number of conditions which will be discussed later in this chapter.

The quantity of heat, q , that will be conducted in time, t , from the hotter to the colder isothermal surface is defined by:

$$q = K \frac{T_1 - T_2}{Z} At \quad (2)$$

where K is a constant known as the thermal conductivity and represents the quantity of heat which flows per unit time through a unit volume of the substance with a unit temperature difference between its faces. The limiting value of $(T_1 - T_2)/Z$ or dT/dZ is known as the thermal gradient at any point.

The rate of flow of heat, commonly referred to as the heat flow, is derived by taking the first derivative with respect to time of Equation (2):

$$Q' = \frac{dq}{dt} = K \frac{T_1 - T_2}{Z} \quad (3)$$

If instead of an isothermal surface we consider another, making an angle ϕ with it, both the flux Q (heat flow per unit area) across the surface and the temperature gradient across the normal to such a

surface will be diminished - the factor being $\cos \phi$ - such that in general the flux across any surface is:

$$Q = -K \frac{\partial T}{\partial z} \quad (4)$$

where the derivative is taken in the direction of decreasing temperature.

While the rate of heat transfer in a body is dependent upon the thermal conductivity, the rise in temperature that this heat will produce will vary with the specific heat, c , and the density, ρ , of the body. A further constant, α , which represents the thermal diffusivity is defined by:

$$\alpha = K/\rho c \quad (5)$$

$$\text{or} \quad \alpha = K/C \quad (6)$$

where C is the volumetric heat capacity and is equal to the product of the density and the specific heat. The importance of the thermal diffusivity will become evident in a later section.

In three dimensions, the general equation for heat conduction in a body (derived by Ingersoll et al., 1954) is defined by Fourier's equation:

$$\frac{\partial T}{\partial t} = \alpha \left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2} \right) \quad (7)$$

$$\text{or} \quad \frac{\partial T}{\partial t} = \alpha \nabla^2 T \quad (8)$$

The solution of any particular problem involving heat flow must first satisfy this equation, either as it stands or in a modified form. Gold and Lachenbruch (1973) state that the main features of the temperature regime in permafrost can be understood in terms of four rather distinct heat-transfer problems:

- 1) The steady one-dimensional flow of heat from the earth's

interior to its isothermal surface, and the depth to the base of the permafrost (0°C isotherm).

- 2) The periodic seasonal variation of surface temperature and the depth to the top of the permafrost (depth of zero annual amplitude variation.
- 3) Long-term variations in surface temperature, the evolution of permafrost and the effects of climatic change, and
- 4) Lateral variations in surface temperature and the effects of bodies of water, variable surface cover and topography.

Details concerning the first and third problems will be addressed in the following section and the next chapter, respectively. The second and fourth problems would be important in small localized studies where measureable factors such as snow cover, vegetation and the daily fluctuations in surface temperature could be incorporated into the analysis. On a regional scale, these factors must be averaged and, in general, are incorporated into the estimate of mean annual ground temperature (MAGT).

4.2 Steady-State Heat Flow

A body in which heat is flowing is said to have reached Steady-state when the temperatures of its different parts do not change with time. Such a state occurs in practice only after heat has been flowing for a long time and no perturbations of surface conditions have occurred. At steady-state, each part of the body loses on one side as much heat as it gains on the other. The temperature, therefore, is independent of the time, t , although it varies from point to point in the body, and is a function of the coordinates x , y and z . Under steady-state conditions, Fourier's

conduction equation⁰ becomes:

$$\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2} = 0 \quad (9)$$

In one dimension, which would include the case of steady flow of heat in any body such that the isothermal surfaces are parallel planes, the general equation reduces to:

$$\frac{d^2 T}{dz^2} = 0 \quad (10)$$

the ordinary derivative being written instead of the partial. The solution of this equation is:

$$T = Bx + C \quad (11)$$

where the constants B and C are determined from the boundary conditions, which are the temperatures T1 and T2 at the parallel plane faces. Thus equation (11) becomes:

$$Q = K \frac{T_1 - T_2}{Z} = K G \quad (12)$$

where Z is the distance between the parallel planes and G is the geothermal gradient.

If it is assumed that the base of the permafrost occurs where the temperature T1 is 0°C, then the depth to the base of the permafrost is given by:

$$Z = K_f (-T_2) / Q \quad (13)$$

where K_f is the thermal conductivity of the frozen material (permafrost), T2 is the long-term mean surface temperature (in °C) obtained by upward extrapolation of the steady-state temperature gradient, and Q is the steady-state heat flow. Low heat flow, low surface temperature T2 and high thermal conductivity favour deep permafrost, Z.

Through the use of this equilibrium equation, the depth to the 0°C isotherm was estimated at various locations where borehole temperature logs were available (Canadian Geothermal Data Collections, 1977, 1979, 1981, 1982).

A further use of this steady-state relationship is in multi-layer problems. As will be discussed in the following section, thermal conductivity of a material is dependent upon the thermal conductivities of its components - water being of prime importance. When water freezes, its conductivity becomes approximately four times as great, hence at the base of the permafrost there will be a marked decrease in thermal conductivity. This condition results in a two-layer problem where the upper frozen layer has a conductivity K_f and a thermal gradient G_f and the lower unfrozen layer has a conductivity K_u and a thermal gradient G_u . Because the same amount of heat Q must be flowing across the boundary:

$$Q = K_f G_f = K_u G_u \quad (14)$$

and the gradient in each layer varies inversely with its conductivity and directly with the regional heat flow.

One application of equation (14) involving the gradient contrast at the base of the permafrost provides an estimate of porosity in terms of ice content. Above the base of the permafrost, interstitial moisture is predominantly in the form of ice (particularly if the depth to the base of the ice-bearing permafrost is considered). Below, moisture is probably in the form of liquid water. At steady-state, equation (14) represents this condition. If the only significant difference between the two layer materials is their moisture state, then the conductivity contrast is a consequence of the fact that ice is approximately four times as conductive as water. Assuming saturation conditions in both layers,

the magnitude of the resulting gradient contrast is a function of the fraction of ice or water present, and is equal to the porosity. This fact is used to provide porosity estimates for the various sequences.

Lachenbruch et al. (1982) present an expression for the porosity, N , in terms of the conductivity contrast of a material in the frozen and unfrozen states. The frozen conductivity, K_f , is expressed as the geometric mean of its constituents (refer to the following section on thermal conductivity) and is defined by:

$$K_f = K_i^N K_g^{1-N} \quad (15)$$

where K_i is the conductivity of ice (approximately $2.3 \text{ Wm}^{-1}\text{K}^{-1}$) and K_g is the geometric mean conductivity of the material grains. The values for conductivity were evaluated at -5°C by Lachenbruch et al. (1982) as they assumed this to be a representative temperature for the frozen permafrost section. Similarly, the conductivity of the thawed material is given by:

$$K_u = K_w^N (\gamma K_g)^{1-N} \quad (16)$$

where K_w is the conductivity of water ($0.56 \text{ Wm}^{-1}\text{K}^{-1}$) at a representative temperature of $+2^\circ\text{C}$. The value of γ - which is approximately 0.983 - adjusts K_g measured at the frozen temperature (-5°C) to the appropriate value at $+2^\circ\text{C}$.

Dividing (15) by (16), replacing the conductivity ratio by the gradient ratio with (14), solving for N , and inserting the numerical values yields:

$$N = 0.772 (\ln G_u/G_f - 0.017) \quad (17)$$

Equation (17) was used in the present study to estimate the porosities of the sequences where discrete frozen and unfrozen

layers could be identified. Again, it should be noted that the major assumption is that the materials are uniform except for the state of their interstitial moisture, and that the amount of unfrozen water in the frozen section is negligible.

A second point to mention is that the value γ of proposed by Lachenbruch et al. (1982) was taken to represent the conversion factor for the grain conductivity variation in the frozen and unfrozen states.

Where available, gradient contrasts for particular sequences were measured and used to calculate the porosity, N , for the sequence. These values are presented in Table 2. An average porosity of 37.6% based on four values was obtained for the Reindeer Sequence, and a value of 18.6% based on a single value was obtained for the Richards Sequence. A typical porosity for a sandstone is estimated at 30%, hence the value calculated for the Reindeer Sequence is probably a representative value. However, the value calculated for the Richards Sequence appears to be a bit high in comparison to a porosity of 6%, typical of mudstones and shales. This may be a reflection of the fact that the Richards Sequence contains a large percentage of unfrozen water in the frozen state therefore making the porosity estimate unreliable. Despite these slight discrepancies, the estimates of porosity calculated using gradient contrasts have been chosen to represent the porosities of the sandstone sequences ($N=38\%$) and the mudstone sequences ($N=19\%$). In fact, slight discrepancies in porosity do not alter the calculated values of heat capacity appreciably.

4.3 Thermal Properties

4.3.1 Thermal Conductivity

While heat flows as a result of a temperature difference and the amount increases in proportion to the difference, the rate of heat transfer depends on the thermal conductivity, K , of the material through which the heat flows. Materials - in this case rocks and sediments in the subsurface - which have high thermal conductivities will be better conductors of heat and, therefore, will transmit heat at a faster rate. Alternatively, low thermal conductivity materials will conduct heat at a slower rate.

An experimental arrangement known as the divided bar apparatus is used for determination of K in soils and rocks (Birch, 1950). In its simplest form the apparatus involves placing a sample between plates of a uniform solid whose conductivity is accurately known. Constant temperatures T_1 and T_2 are established on either side of the sample and after a short while steady-state is attained. The flow of heat, Q , through the sample is necessarily the same as through the standard plates. If the thicknesses of the sample and the plates are the same, then, the conductivity of the unknown material is readily calculable.

An alternative device (Jackson and Taylor, 1965) used with soils is the line heat source probe which is inserted into a soil or sediment sample. The probe is basically a length of steel tubing inside which are a heating coil and a temperature measuring device. A known quantity of heat is supplied by a monitored electric current through the coil. The rate of dissipation of the heat depends on the conductivity of the soil, and is shown by the rate of temperature rise of the probe.

Various authors have discussed models for the conduction of heat in granular materials so as to predict the thermal conductivity on the basis of the composition of the material. A comprehensive study by Johansen (1972, 1973, 1979) led to the conclusion that the conductivity of saturated materials is fairly well represented by the geometric mean equation:

$$K = K_1^{V_1} K_2^{V_2} K_3^{V_3} \dots \quad (18)$$

where K is the conductivity of the soil or sediment; K_1 , K_2 , K_3 are the conductivities of the constituent materials - including water and ice - and V_1 , V_2 , V_3 are the volume fractions of each of the constituents. The equation gives large errors when there is a large difference between the lowest and the highest conductivity; hence the air present, with its very low conductivity, makes the equation unsuitable for unsaturated soils.

Equation (18), however, illustrates the various components which contribute to the overall conductivity of a material. For example, the effect of ice on thermal conductivity becomes apparent when compared to the effect of water.

An alternative method for estimating the thermal conductivity involves the use of thermal gradient and regional heat flow. This method was employed in the present study.

Using all of the available borehole temperature logs (Canadian Geothermal Data Collections for Northern Wells) in conjunction with the borehole lithologies (Figures 7a through 7g) an average thermal gradient was estimated, where possible, in the frozen and unfrozen layers. These values have been grouped according to the sequence for which they apply and are reported in Table 3.

Also given in Table 3 are the values of thermal conductivity for an assumed constant and uniform heat flow of 60 mWm^{-2} . The

assumption that Q remains constant and uniform is supported by the consistency of heat flow values calculated at various locations along the Alaskan arctic coast (Lachenbruch et al., 1982) and the uniformity of deep temperature gradients across the Mackenzie Delta region (Judge, 1986). Judge (personal communication) suggests that 60 mWm^{-2} is a good estimate of the regional heat flow. Lachenbruch et al. (1982) obtained an average of 55 mWm^{-2} for the Prudhoe Bay, Alaska.

The conductivity estimates reported in Table 3 were then averaged to give a frozen and unfrozen conductivity for the various sequences. The resulting values are provided in Table 4. These values will be employed in the thermal modelling of permafrost aggradation in the Mackenzie Delta region. More details concerning thermal modelling are provided in Chapters 6 and 7.

4.3.2 Heat Capacity

The volumetric heat capacity, C , is the quantity of heat (in Joules) which raises or lowers the temperature of 1 m^3 of material by 1°C . As such, it determines the amount by which the temperature of a body will change in response to heat being transferred through it. In heterogeneous materials such as sediment, the heat capacity is simply the sum of the heat capacities of the proportional parts and is defined by:

$$C = C_s V_s + C_w V_w + C_i V_i \quad (19)$$

where V_s , V_w and V_i are the volume fractions of solid, water and ice, respectively and C_s , C_w and C_i are the volumetric heat capacities. The heat capacity of air may be ignored because it is extremely low. At 20°C , $C_w = 4.18 \times 10^8 \text{ Jm}^{-3} \text{C}^{-1}$ and $C_i = 1.83 \times 10^8 \text{ Jm}^{-3} \text{C}^{-1}$. The heat capacity of the water content

is the largest component of the heat capacity of the material.

Therefore, it is important that an estimate of the porosity be made.

Using equation (19) for saturated frozen and unfrozen sandstones (N=38%) and for saturated frozen and unfrozen mudstones (N=19%), values for the volumetric heat capacity were calculated and are shown in Table 5. The heat capacities of the constituent materials were obtained from Williams (Table 5.2b, 1982).

4.3.3 Thermal Diffusivity

If a material initially at a uniform temperature is subjected to a temperature disturbance, there will be a flow of heat to or from the surface. Change of temperature within the body occurring as a result of the applied surface temperature will clearly depend upon the ease with which heat flows in that material, namely the thermal conductivity, and upon the temperature change associated with the gain or loss of heat by that material, namely the heat capacity. This dual dependence is reflected in a quantity known as the thermal diffusivity, α , which is simply the ratio of the thermal conductivity to the heat capacity:

$$\alpha = K/C \quad (20)$$

This property is of importance when describing the transient response of a material to a temperature disturbance and appears in the general form of Fourier's conduction equation (6).

This equation implies that the rate of change of temperature with time is equal to the diffusivity times the rate of change of temperature gradient with depth. Thus, if the temperature gradient in uniform ground varies with depth - such as the case in which the thermal gradient is still responding to a surface temperature disturbance and has not yet attained equilibrium - temperatures are changing at a rate dependent on the thermal diffusivity.

Using the values of conductivity calculated for each sequence and the estimates of heat capacity, thermal diffusivities were calculated for each of the predominant sequences. These values are reported in Table 6.

Variations in diffusivity were incorporated into the thermal models and results indicate that these variations have little to no effect on the final permafrost thickness. For this reason, an average diffusivity of $0.017 \text{ cm}^2 \text{ s}^{-1}$ was used for all sequences and was incorporated into the thermal models.

4.4 Subdivision of Region

On the basis of variations in the subsurface lithologies, the Mackenzie Delta was subdivided into five sub-regions, each having similar subsurface geological characteristics (Figure 10). These include:

- 1) The Kugmallit Bay region, which encompasses the northeastern portion of Richards Island near Kugmallit Bay and is characteristic of the old Mackenzie Delta. The subsurface is dominated by thick Iperk Sequence sediments underlain by a thick accumulation of Kugmallit Bay sands.
- 2) The Tuktoyaktuk Peninsula region, which extends from Kugmallit Bay on the west and the tip of the peninsula to the northeast. Here, the subsurface is dominated by relatively thick Iperk sediments underlain by Reindeer sediments.
- 3) The western portion of Richards Island which has relatively thin Iperk sediments underlain by thick Richards Sequence mudstones.

These three onshore regions have been exposed to similar surface temperature conditions during the Late Pleistocene and Holocene as will be seen in the following chapter on Surface Temperature History. The fourth and fifth sub-regions have been exposed to more unique surface temperature conditions and for this reason are separated into two additional sub-regions.

- 4) The modern Mackenzie Delta where surface temperature conditions are constantly changing in response to migrating river channels, is characterized by relatively thin Iperk sediments underlain by thick Reindeer Sequence sediments. This sub-region will be modelled according to the surface temperature conditions outlined in the following chapter.
- 5) The offshore Beaufort Shelf region has a similar lithologic character to the Kugmallit Bay region along its southern edge but which gradually changes to Iperk Sequence sediments in the offshore. Of course, as mentioned earlier, the Iperk Sequence sediments of the upper sand-gravel unit thin towards the far offshore, hence changing the composition of the deeper sediments. However, this region will be modelled as though lithology stays constant with the length of time of submergence changing.

Each of these five sub-regions was then assigned a value of thermal conductivity which would represent the dominant subsurface lithologies (Table 7). In those regions where the upper sequence was generally thick, a thermal conductivity corresponding to the Iperk Sequence sediments was assigned. Alternatively, where Iperk sediments were thin, the deeper sequence thermal conductivity was assigned. In this way, the subsurface regime could be reduced to a simple single layer and, therefore, could be readily solved. Table 7

also reports the values of porosity assigned to each sequence and indicates whether the values apply to frozen or unfrozen sediments.

Although five sub-regions have been identified, there are other unique subsurface and surface environments which merit separate discussions, for example, the Mackenzie Trough area in the western Beaufort Sea. This region has had a complex surface temperature history as a result of repeated glaciation and sediment accumulation and, therefore, should be modelled separately from the Beaufort Shelf region. These complex environments are simply noted in this thesis and will not be modelled, although a slightly more complex paleoclimate may be inferred from the paleoclimatic history presented later in this study.

TABLE 2

SITE	THERMAL GRADIENT		CALCULATED POROSITY	SEQUENCE
	(°C km ⁻¹) (f)	(u)		
29	16	28	.392	Reindeer
27	16	29	.417	Reindeer
47	18	26	.253	Reindeer
85	16	30	.441	Reindeer
50	19	25	.186	Richards

Average porosity for Reindeer = 0.376 = 37.6%

Average porosity for Richards = 0.186 = 18.6%

Table 2: Porosity estimates for the various sequences based on gradient contrasts at the base of the permafrost.

TABLE 3a
IPERK SEQUENCE

SITE	DEPTH RANGE (m)	GRADIENT (°C m ⁻¹)	THERMAL CONDUCTIVITY (W. m ⁻¹ K ⁻¹)
91	100-300	.0127	(f) 4.72
121	100-200	.0200	(f) 3.00
121	200-300	.0094	(f) 6.38
64	225-600	.0078	(f) 7.69
78	0-147	.0177	(f) 3.89
42	0-300	.0183	(f) 3.27
68	0-350	.0147	(f) 4.09
66	0-275	.0103	(f) 5.82
50	0-350	.0193	(f) 3.11
51	0-125	.0088	(f) 6.82
37	0-400	.0095	(f) 6.32
33	0-250	.0240	(f) 2.50
16	0-275	.0142	(f) 4.24
31	0-135	.0293	(f) 2.04
29	0-150	.0161	(f) 3.72
46	0-150	.0098	(f) 6.13
118	0-200	.0160	(f) 3.75
117	200-300	.0137	(f) 4.39
100	0-260	.0144	(f) 4.18

u-unfrozen, f=frozen

Table 3a: Estimates of thermal gradient and
thermal conductivity for Iperk
Sequence sediments

TABLE 3b
KUGMALLIT SEQUENCE

SITE	DEPTH RANGE (m)	GRADIENT (°C m ⁻¹)	THERMAL CONDUCTIVITY (W m ⁻¹ K ⁻¹)	
66	275-500	.0103	(f)	5.82
16	275-480	.0142	(f)	4.24
118	200-500	.0160	(f)	3.75
99	500-600	.0087	(f)	6.92
101	500-700	.0250	(f)	2.40

u-unfrozen, f=frozen

Table 3b: Estimates of thermal gradient and
thermal conductivity for Kugmallit
Sequence sediments

TABLE 3c
RICHARDS SEQUENCE

SITE	DEPTH RANGE (m)	GRADIENT (°C m ⁻¹)	THERMAL CONDUCTIVITY (W m ⁻¹ K ⁻¹)
107	500-750	.0280	(u) 2.14
75	200-400	.0214	(u) 2.81
68	350-600	.0272	(u) 2.20
74	375-750	.0181	(u) 3.31
50	400-600	.0250	(u) 2.40
77	200-700	.0240	(u) 2.50
80	225-450	.0200	(u) 3.00
80	0-225	.0488	(u) 1.23

u=unfrozen, f=frozen

Table 3c: Estimates of thermal gradient and thermal conductivity for Richards Sequence sediments

TABLE 3d
REINDEER SEQUENCE

SITE	DEPTH RANGE (m)	GRADIENT (°C m ⁻¹)	THERMAL CONDUCTIVITY (W m ⁻¹ K ⁻¹)
85	300-400	.0160	(f) 3.75
85	400-525	.0300	(u) 2.00
19	100-175	.0560	(u) 1.07
42	300-500	.0260	(u) 2.31
37	400-600	.0195	(u) 3.08
33	285-700	.0301	(u) 1.99
16	485-600	.0416	(u) 1.44
31	135-275	.0144	(f) 4.17
29	150-335	.0161	(f) 3.72
29	400-700	.0283	(u) 2.12
46	370-450	.0310	(u) 1.94
47	75-300	.0176	(f) 3.42
47	375-525	.0260	(u) 2.30
85	150-300	.0260	(f) 2.31
85	300-400	.0160	(f) 3.75
85	400-525	.0300	(u) 1.99
117	300-500	.0137	(f) 4.39
28	145-300	.0241	(f) 2.06
28	225-500	.0272	(f) 2.21
23	150-250	.0250	(f) 2.40
48	150-300	.0127	(f) 4.74
48	300-375	.0233	(u) 2.57
27	0-335	.0163	(f) 3.67
27	370-550	.0289	(u) 2.08

u=unfrozen, f=frozen

Table 3d: Estimates of thermal gradient and thermal conductivity for Reindeer Sequence sediments

TABLE 4

SEQUENCE	CONDUCTIVITY ($W\ m^{-1}\ K^{-1}$)		LITHOLOGY
	mean	s.d.	
IPERK	4.5	1.6	(f) sandstone
KUGMALLIT	4.6	1.8	(f) sandstone
REINDEER	3.4	0.9	(f) sandstone
	2.1	0.5	(u)
RICHARDS	2.4*	0.6	(u) mudstone

*Conductivity for frozen material will be similar due to low ice content.
u-unfrozen, f-frozen

Table 4. Average thermal conductivities and standard deviations for the various sequences

TABLE 5

SANDSTONE (Iperk, Kugmallit, Reindeer)	$C_f = (2.2 \times 10^6)(.62) + (1.83 \times 10^6)(.38)$ $= 2.089 \times 10^6 \text{ Jm}^{-3} \text{ } ^\circ\text{C}^{-1}$
	$C_u = (2.2 \times 10^6)(.62) + (4.18 \times 10^6)(.38)$ $= 2.794 \times 10^6$
MUDSTONE (Richards)	$C_f = (1.6 \times 10^6)(.81) + (1.83 \times 10^6)(.19)$ $= 1.614 \times 10^6$
	$C_u = (1.6 \times 10^6)(.81) + (4.18 \times 10^6)(.19)$ $= 1.755 \times 10^6$

Table 5: Calculated heat capacities for sandstone
and mudstone dominant lithologies

TABLE 6

IPERK (f)	$\alpha = 4.5 / (2.1 \times 10^6) = .021 \text{cm}^2 \text{s}^{-1}$
KUGMALLIT (f)	$\alpha = 4.6 / (2.9 \times 10^6) = .022$
REINDEER (f)	$\alpha = 3.4 / (2.1 \times 10^6) = .016$
(u)	$\alpha = 2.1 / (2.1 \times 10^6) = .010$
RICHARDS (u)	$\alpha = 2.4 / (1.6 \times 10^6) = .015$

Average Diffusivity = $0.017 \text{cm}^2 \text{s}^{-1}$
u-unfrozen, f=frozen

Table 6: Calculated thermal diffusivities for the various sequences

FIGURE 10 Map Showing Subdivision of Region

TABLE 7

SUB-REGION	DOMINANT SEQUENCE	CONDUCTIVITY ($\text{W m}^{-1} \text{K}^{-1}$)	POROSITY
KUGMALLIT BAY	IPERK	4.5	38%
TUKTOYAKTUK PENINSULA	REINDEER	3.4	38%
WESTERN RICHARDS ISLAND	RICHARDS	2.1	19%
MODERN MACKENZIE DELTA	REINDEER	3.4	38%
OFFSHORE BEAUFORT SHELF	IPERK	4.5	38%

Regional Heat Flow = 60 mW m^{-2}

Thermal Diffusivity = $0.017 \text{ cm}^2 \text{ s}^{-1}$

Table 7: Modelling parameters for the various
sub-regions

CHAPTER 5 - SURFACE TEMPERATURE HISTORY

The subsurface thermal regime of any region is characterized by a number of parameters. These include the regional heat flow and the thermal properties of the subsurface - as discussed in the previous chapter - as well as the surface temperature and its fluctuation through time.

The aggradation of permafrost is controlled by variations in the surface temperature which depend upon climatic conditions and energy exchanges between the ground surface and the atmosphere. Because the exchange of heat at the ground surface involves so many processes which are controlled by the nature of the ground surface - for example, relief, exposure, vegetation - it is difficult to assess the individual effect of each on the permafrost regime. This is particularly true of areas where surface temperatures are close to 0°C . In these areas, slight variations in the mean annual surface temperature of a degree or more over small distances dictate whether permafrost will be present or not.

Where the mean temperatures are, in general, several degrees below 0°C (such as the Mackenzie Delta region), only quite extreme surface conditions will result in the absence of permafrost. Large bodies of water, such as the Mackenzie River, are an example. If the water does not freeze to the bottom in the winter, the ground surface - that is, the river bed - is never frozen. If the water body is wide enough, the permafrost on either side will not meet at any depth under the water body. Under these conditions, there will be a hole in the otherwise "continuous" permafrost which is a direct result of an increased local mean annual surface temperature.

Because the Mackenzie Delta is for the most part underlain by

continuous permafrost except beneath major rivers and lakes and newly deposited unconsolidated sediments, a large-scale examination of the effects of long-term surface temperature fluctuations is best made. This may be explained in terms of the following equations.

If one considers the effect of imposing a cyclic temperature wave on a uniform solid body (Williams, 1982), then the amplitude S at depth Z below the surface is defined by:

$$S_z = S_b \exp(-Z/\pi/\alpha t) \quad (21)$$

where S_b is the amplitude at the surface, α is the thermal diffusivity and t is the wave period. The effect of diffusivity on the wave amplitude should be noted.

The delay, at depth Z below the surface, between the occurrence of a certain point on the temperature wave (for example, the summer maximum temperature) and the surface boundary is defined by:

$$\Delta t = \frac{Z}{2} \frac{\sqrt{t}}{\sqrt{\alpha\pi}} \quad (22)$$

This equation indicates that as the depth increases, the delay Δt becomes more marked. Hence, there exists a time lag between the subsurface temperatures and their associated surface temperatures which increases in proportion to the depth. A further equation:

$$Z_1 = 2/\pi\alpha t \quad (23)$$

gives the approximate depth Z_1 at which a cyclic temperature fluctuation of period t will be felt. This equation clearly indicates that temperature fluctuations of long periods will be felt at much greater depths than the shorter period ones.

Due to the relatively low thermal diffusivity of rocks, any surface temperature changes which occur at the ground surface propagate slowly downward, and remain recorded but attenuated and dispersed at depth under the surface (Cermak, 1970). For example, if

subsurface materials have an average diffusivity of $0.017 \text{ cm}^2 \text{ s}^{-1}$, then in order for a surface temperature disturbance to be felt at 500m (depth of permafrost), the period of the temperature disturbance should be no less than approximately 370 years. This indicates that in order to alter the permafrost thickness through variations in the surface temperature, long-term fluctuations should be analysed.

An understanding of the present thermal regime in terms of the observed permafrost thicknesses in the Mackenzie Delta region requires consideration of the climatic trends for many tens of thousands of years. In general, there are four factors which have had in the past or which have at present a significant effect on the permafrost regime. These include air temperature variations, the glacial history, - in terms of the extent and duration of glacial ice cover - fluctuations in sea level associated with the glacial periods, and the presence or absence of major bodies of water such as the Mackenzie River. The effect of each of these four factors on the surface temperature history will be discussed in the following pages so that a paleoclimatic model may be clearly defined.

5.1 Air Temperature

Air temperature is probably the most important factor to affect the surface temperature history. In the absence of such varying conditions as snow cover, vegetation or relief, air temperature is generally a couple of degrees lower than the actual ground surface temperature. In a tundra region, such as the Mackenzie Delta, the different conditions at the ground surface (excluding the existence of bodies of water) are generally consistent throughout the region. Therefore, an average ground surface temperature may be assigned on

the basis of air temperature alone. This assumption is made in the present study for the purposes of reconstructing a paleoclimate for the Mackenzie Delta region.

Estimates of air temperatures and their fluctuation throughout the Late Pleistocene and Holocene - a period of approximately 100,000 years - are based on air temperature information derived from 1) oxygen isotope records for the Greenland ice cap (Dansgaard et al., 1971), and 2) amino acid temperature estimates for the Alaskan Beaufort Shelf (Brigham and Miller, 1983; Smith, 1986). Associated with each of these methods are a number of problems which will be discussed briefly in the following two sections.

5.2 Oxygen Isotopes

The use of oxygen isotopes to estimate paleoclimates requires some understanding of the characteristics of stable isotopes in the environment.

The element oxygen has three stable isotopes ^{16}O , ^{17}O and ^{18}O which are present in the environment in varying amounts according to the energy level in the phase in which they occur. Heavier isotopes are preferentially incorporated into the lower energy phases, hence, the ratio of ^{18}O to ^{16}O for ice (solid phase of water) will be higher than the $^{18}\text{O}/^{16}\text{O}$ for water vapour.

It is customary to report relative changes in ratios of isotopes in terms of 'per mille' variations. Therefore,

$$\delta_x^{18\text{O}/16\text{O}} = \frac{R_x - R_{\text{standard}}}{R_{\text{standard}}} \times 1000 \quad (24)$$

or, equivalently,

$$\delta^{18}\text{O} = \frac{R_x - R_{\text{standard}}}{R_{\text{standard}}} \times 1000 \quad (25)$$

where R_x is the ratio of ^{18}O to ^{16}O in the sample and R_{standard} is the ratio of ^{18}O to ^{16}O in a standard. The present standard ratio used is Standard Mean Ocean Water (SMOW) which would, according to Equation (25) give a 'per mille' variation of $\delta^{18}\text{O} = 0\text{‰}$.

In order to gain more information concerning the conditions of an environment, samples may be collected and analysed to determine their $\delta^{18}\text{O}$ value. In general, these values are temperature dependent and conform to the following conditions:

- 1) The fractionation constants change with temperature.

(Fractionation or the partitioning of elements in nature results from the difference in mass between isotopes. The fractionation constant defines the equilibrium concentration between substances).

- 2) A particular isotopic ratio (such as $\delta^{18}\text{O}$ for water) will increase with temperature. This relationship is defined by:

$$\delta^{18}\text{O} = 0.695t_a - 13.6 \quad (26)$$

(after Dansgaard et al., 1964), where t_a is the mean annual air temperature.

It is this second condition which is of importance in relating variations of temperature to varying isotopic ratios.

Dansgaard et al. (1971) present an $\delta^{18}\text{O}$ profile for an ice core retrieved from the Camp Century Ice Cap in Greenland (Figure 11). Variations in $\delta^{18}\text{O}$ with depth are clearly identified and are

2

of/de

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MicroD

believed to correspond to changes in air temperature at the time of mass accumulation on the glacier. However, differences in altitude and location of snow deposition (continental effect) make it difficult to assign a set temperature.

The less negative the 'per mille' variations, the higher the concentration of the heavier isotope ^{18}O in the sample. Accordingly, this corresponds to a higher air temperature at the time of deposition. The amplitude variations steadily decrease with depth into the core due to diffusion because these greater core depths correspond to older deposit ages.

It has been estimated (Dansgaard et al., 1971) that a change in $\delta^{18}\text{O}$ of +1‰ is associated with a rise in air temperature of approximately 15°C (refer to Equation (26)). This relationship was arrived at by taking into account the effect of altitude variations of the glacier, which would contribute to a 3‰ increase.

Using the 'per mille' / temperature relationship described above, average mean annual ground temperatures were estimated for the Sangamonian Interglacial and the Early and Mid-Wisconsinan glacial periods relative to present mean annual ground temperatures. Because we are dealing with relative values and the Holocene MAGT is known, direct estimates of the Pleistocene MAGTs may be made.

5.3 Holocene Mean Annual Ground Temperatures

In order to simplify the thermal modelling, it is assumed that Holocene air temperatures have remained relatively constant in the onshore regions of the Mackenzie Delta. Any small, short-term temperature fluctuations are expected to have a negligible effect on the permafrost base. Mean annual ground temperature estimates for the Mackenzie Delta region are shown in Figure 2 (after Young and

Judge, 1985).

The mean annual ground temperatures at various locations were determined using: a) upward extrapolation of the geothermal gradient for those sites where equilibrium thermal profiles were known and where no recent climatic changes had taken place, and b) averaging mean annual ground temperature measurements over at least a twelve month period - obtained from Atmospheric Environment Service (AES) ground temperature stations and Geological Survey of Canada (GSC) data from short temperature cables.

An approximate mean annual ground temperature for the Mackenzie Delta region based on values for emergent areas is estimated at -9°C . This average does not incorporate measurements obtained in thermally disturbed environments such as the modern Mackenzie Delta - where surface conditions and temperatures reflect varying states of emergence and submergence.

In the modern Mackenzie Delta, a mean annual ground temperature was assigned on the basis of mean annual bottom water temperatures for Shallow Bay. An average value of $+1^{\circ}\text{C}$ (Burgess, 1977) was chosen to represent surface conditions in the modern Mackenzie Delta as this area is, in general, submerged beneath warm Mackenzie River outflow.

5.4 Amino Acid Racemization Dating

A second source of Pleistocene air temperature estimates for the Mackenzie Delta region is derived from Amino Acid Racemization Dating (Geochronology). This technique is used in Pleistocene stratigraphic studies as a tool for relative age dating and correlation of equivalent strata or for the absolute dating of

deposits.

The method is based upon the detection of changes in the amino acid isomer distributions that accompany fossilization. A number of fossil materials including bone, have been employed with this method (Bada, 1972). As well, correlation and relative age dating of marine, freshwater and terrestrial molluscs have been carried out (Brigham, 1982, 1983).

The method is particularly useful for the correlation and relative age dating of equivalent strata which have experienced similar temperature histories and diagenetic conditions.

5.4.1 Analytical Methods

Most amino acids - the building blocks of proteins - may occur in one or two optically active forms, or enantiomers, which are mirror images of each other. Those forms that rotate polarized light to the right are D-amino acids (dextrorotatory) and to the left are L-amino acids (levorotatory).

Constituent amino acids that comprise the proteins within the carbonate matrix of living molluscs are exclusively of the L-stereoisomer configuration. After the death of the organism, the L-amino acids reversibly convert to the D-configuration until an equilibrium mixture is established. This process is referred to as "Racemization". Hence, fossil shells of increasing age contain a greater proportion of the D-amino acids and the ratio of D/L amino acids expresses the extent of diagenesis from the L- to the D-configuration.

This process (racemization), occurs slowly at ambient temperatures but is greatly accelerated by elevated temperatures or the presence of acids or alkalis in the system. Because the L- to D-

conversion is time dependent, the D/L ratio of a given amino acid may be used as an indicator of fossil age. For samples from high latitudes, complete racemization may not be achieved until several millions of years after death because of the cold climate. It should also be noted that the rate of racemization, and hence the age range, varies with the amino acid being used.

5.4.2 Paleotemperature Estimates

Temperature is the most critical rate-controlling factor governing protein diagenesis. In general, a 4°C temperature rise will approximately double the rate of racemization from the L- to D-configuration. However, for relatively young samples (for example, 125,000 years B.P.) at temperatures lower than about -10°C the rate of racemization is so slow that the change in the absolute value of the D/L ratio, created by fluctuations in temperature of a few degrees, is very small. Hence the method cannot detect temperature differences of $\pm 2^\circ\text{C}$ in this time/temperature range (Brigham and Miller, 1983).

Using the relationship between the extent of racemization, absolute time and diagenetic temperature (reviewed by Schroeder and Bada, 1976) the effective diagenetic temperature (EDT) - that is, the temperature associated with an effective rate constant estimated by integration through a probable temperature history for a sample - is defined by:

$$T = \frac{6131}{16.33 - \log \left\{ \frac{\ln \left(\frac{1 + D/L}{1 - .77D/L} \right) - 0.0195}{1.77t} \right\}} \quad (27)$$

where t is the time since the death of the organism (Brigham and

Miller, 1983).

Paleotemperature estimates derived from Equation (27) are most accurate when the absolute age (t) of a sample is known. Brigham and Miller (1983) present various paleoclimatic reconstructions for the Alaskan Arctic Coastal Plain based on D/L ratios of molluscs from Sangamonian Interglacial deposits. Using the paleoclimatic record as a guide for interglacial periods of "known" temperature (EDT estimates), models were developed to derive appropriate EDT's for periods lacking in data, especially full-glacial episodes. They conclude that the climate of the North Slope (Alaskan Coast) during the glacial episode was at least 8°C colder than today and that glacial ground temperatures probably averaged lower than -18°C .

Based on the degree of racemization in fossil foraminifers Smith (1986) presents a thermal model which reflects postulated conditions during alternating periods of exposure and inundation for various sites in the Canning River Delta in northern Alaska. The model incorporated both the probable sea level and paleoclimatic history of the area and was then used to determine possible ages from the raw amino acid ratios. The model was then shown to provide the best fit to the environmental conditions implied by microfauna. Figure 12 (after Smith, 1986) shows the model results for the paleoclimatic history of the Beaufort Sea continental shelf.

5.5 Glacial History

The glacial extent and duration in the Mackenzie Delta during the Wisconsin glacial period remains questionable despite efforts made by various researchers to define conclusively the limits of glaciation in the region.

Extensive research on the glacial history has been carried out

in the Mackenzie Delta and the adjacent Yukon Coastal Plain by Mackay (1963) and Rampton (1988). The extent of glacial erosion described by these researchers suggests that the Late Pleistocene ice sheets were largely wet-base ice masses since large-scale erosion and deposition seem to require a sliding bed where freeze-thaw processes can occur. Therefore, one might expect the glaciated areas to have experienced a relative warming at their surface as a consequence of the warmer temperatures (approximately 0°C - the pressure-melting point) at the base of the ice sheet.

In response to warmer ground surface temperatures produced by the glaciers, temperature disturbances would migrate down into the ground and eventually cause degradation at the base of the permafrost. Hence, permafrost, if it were present prior to glaciation, would be expected to have undergone some degradation during the glaciation.

5.6 Late Pleistocene Glaciations

The greater part of the Mackenzie Delta was glaciated during the Wisconsinan. The ice did not advance westward beyond the front of the Richardson Mountains and was probably absent from the northeastern tip of the Tuktoyaktuk Peninsula. The map shown in Figure 13 (after Rampton, 1986) best represents the knowledge to date acquired concerning the glacial limits in the Mackenzie Delta region.

The maximum extent of Early Wisconsinan ice has been termed the Buckland Glaciation (Rampton, 1982). The limit of this glaciation lies against the northern slope of the Richardson Mountains, along the southern edge of the Yukon Coastal Plain, and

just west of Herschel Island. Offshore, the limit reaches the Mackenzie Trough and heads back toward Richards Island. Field evidence in support of glaciation has been found on Garry and Kendall Islands and in the northern portion of Richards Island. Air photo interpretation suggests that Hooper and Pullen Islands were also glaciated (Mackay, 1963). East of the Mackenzie Delta, the limit lies to the north of the Eskimo Lakes and southern Liverpool Bay, and extends northward into Kugmallit Bay to the west of the town of Tuktoyaktuk. This limit has also been termed the Mason River Limit (Rampton, 1986).

Plant material (14.4ka) and platey silts (14.1ka and 17.9ka) stratigraphically located above tills presumably associated with these limits, provide an upper limit to the latest ice advance. Shells within glacial(?) terraces northwest of Richards Island date at >35,000 and >36,000 years B.P. (Rampton, 1986). Pleistocene sand and ice wedges examined at Hooper island indicate an age younger than Early Wisconsinan. Sediments forming the unglaciated terrain of Garry Island and associated areas were deposited more than 40,000 years B.P. (Mackay and Matthews, 1983). The maximum extent of Early Wisconsinan ice, therefore, must have occurred more than 35,000 years ago.

The age for the Buckland Glaciation agrees with the age determined from the amino acid air temperature history proposed by Smith (1986) which places the Early Wisconsinan glacial period between 60,000 and 75,000 years B.P. The Buckland ice advance is assumed to have had its maximum extent around the middle of this period, namely, 65,000 to 70,000 years B.P.

During deglaciation, a major standstill or readvance may have occurred and constructed a moraine-outwash complex parallel to the

Yukon coast (refer to Figure 13). This readvance or stand still has been named the Sabine Phase of the Buckland Glaciation (Rampton, 1982). Evidence of the Sabine limit is also present in the southern portion of the Tuktoyaktuk Peninsula in a direction parallel to its strike, and at the northern end of the Eskimo Lakes. These two limits on either side of the modern Mackenzie Delta must logically be connected. Therefore, despite the lack of information on the western side of Richards Island, it is likely that a lobe of Sabine ice extended down the modern Delta and into Mackenzie Bay. This ice, however, would have been absent from Richards Island and the adjacent offshore area to the north.

A second major ice advance, termed the Classical or Late Wisconsinan advance is believed not to have extended northwest of Shallow Bay. The inferred date of the maximum extent of this advance is between 13,500 and 14,000 years B.P. (Prest, 1969; Mackay et al., 1972). According to Smith (1986) the Late Wisconsinan occurred between 15,000 and 30,000 years B.P.

The minimal coverage of this ice suggests that it was unlikely to have caused significant alteration of the permafrost regime in the greater part of the Mackenzie Delta. Assuming this to be the case, any noteworthy modifications of the permafrost regime must have occurred prior to the Late Wisconsinan.

Mackay and Matthews (1983) conclude that permafrost must have been present at Hooper Island since at least Early Wisconsinan, otherwise both the wedge ice and the lens ice in the glacially deformed silts would have melted. In addition, it would seem probable that permafrost in excess of 700m thickness (such as that observed beneath the northeastern portion of Richards Island) would

require a growth period - given a sand-dominant, high thermal conductivity lithology, and a low temperature - equal in length to the greater part of the Wisconsinan, if not longer (Mackay, 1972).

Whether or not the glaciations had any significant effect on the present permafrost distribution and thickness in the Mackenzie Delta region remains questionable. However, in order to provide an answer, the effect of the glacial advances in terms of the associated rise in surface temperature of the glaciated areas will be incorporated into the paleoclimatic models. Surface temperature rises to 0°C will be applied to the permafrost regime. For the purposes of this study, the Late Wisconsinan Glaciation is neglected as its area of influence is thought to have been restricted to such a small portion of the delta. This glaciation, however, does help to explain the limited occurrence of permafrost associated with the modern Mackenzie Delta and the Mackenzie Trough.

5.7 Submergence-Emergence

A third factor which is believed to have had a significant effect on the surface temperature history is the submergence-emergence status of the Mackenzie Delta region. During periods of submergence, the mean annual ground temperature would necessarily have increased relative to sub-aerial ground temperatures in response to warmer seabottom temperatures - particularly if submergence occurred during the waning stages of a major ice advance. Alternatively, during periods of maximum glacial extent, sea level would in general have been low and many coastal regions would have been exposed to extreme surface temperatures.

Submarine permafrost, unlike its subaerial counterpart, can aggrade and degrade in a variety of ways. Mackay (1972) classifies

offshore permafrost into two main types: equilibrium and disequilibrium permafrost. The former is said to exist in those areas where the mean annual seabottom temperature is negative, and has remained unchanged over a substantial period of time. The thickness of this equilibrium permafrost depends on the annual seabottom temperature, water salinity, soil properties, the geothermal heat flux and the sedimentation rate (Mackay, 1972, Figure 5a). Judge (1974) estimates that the maximum equilibrium permafrost thickness offshore - away from the shoreline - may range from 25 to 75m in water depths of 250m, depending on whether seabottom temperatures are negative. If a thin veneer of sediment overlays bedrock, permafrost may be as thick as 125m. The effect of nearby shorelines with much lower mean annual seabottom temperatures than the neighbouring seabottom, results in a lateral flow of heat and, consequently, local thickening of the equilibrium offshore permafrost (Judge, 1974).

Disequilibrium permafrost is not in equilibrium with either the seabottom temperature or the heat flux from the earth's interior. As a result, the permafrost must be either aggrading or degrading. Most disequilibrium permafrost is probably relic degrading permafrost.

As coastal regions became inundated by rising sea levels at the end of the last major glaciation - namely, the Late Wisconsinan - permafrost terrain became gradually submerged and subjected to warmer seabottom temperatures. As a result, permafrost which had aggraded to great depths during the Early Wisconsinan - when temperatures were significantly lower than at present - began to degrade. Thick Early Wisconsinan permafrost in the Mackenzie Delta, may provide an explanation for the great thicknesses of permafrost

observed offshore today.

In the shallow waters of the Beaufort Sea, seabottom temperatures are primarily positive due to Mackenzie River outflow, whereas in deeper waters they are negative. As sea level began to rise in the Beaufort Sea region following deglaciation the land surface became submerged; initially in warm nearshore waters and then in water with negative temperatures. Whether or not this nearshore warming phenomenon has persisted throughout all marine transgression episodes, depends upon the temperatures of the Mackenzie River outflow and the glacial outwash waters. For modelling purposes, the initial warming period will be excluded and the average submergence seabottom temperatures will be taken as -1.8°C (after Mackay, 1972).

Hill and Blasco (1986) present a relative sea level curve for the Beaufort Shelf (Figure 14) which was constructed from results of radiocarbon dating of peat and peaty clay samples recovered from offshore geotechnical boreholes. Each radiocarbon date was assessed palynologically for indications of reworking and for environment of deposition. From the analysis and from information gained by relating samples to seismic profiles, the former position of each sample - relative to paleo-sea level - was located to within a few metres.

Resulting data indicate a fairly constant rise in relative sea level (0.55cm yr^{-1}) of 140m in the past 27,000 years. A minor lowering between 21,000 and 8,000 years B.P. is also indicated. The relatively small rise and fall around 15,000 years B.P. may reflect the migration of a peripheral bulge related to the Late Wisconsinan Glaciation (Hill et al., 1985).

The combined effects of basin subsidence due to movement along

major growth faults and sediment accumulation, as well as sediment loading and consolidation subsidence, were estimated by Hill et al. (1985) to account for a total subsidence of 35m. Glacio-eustatic effects at the margin of the Wisconsinan ice sheet, therefore, would account for the remaining 105m of RSL rise since 27,000 years B.P.

The relative sea level curve also indicates that the crust had not yet recovered from the Early Wisconsinan ice advance when the Late Wisconsinan ice moved into the area. This suggests the presence of a Mid-Wisconsinan ice sheet, which may have been a stationary or an ablating remnant of Buckland ice (Hill et al., 1985). This proposed Mid-Wisconsinan episode may be related to the major readvance discussed by Rampton (1982) as the Sabine Phase. If this were the case, the interstadial between the Early and Late advances would have been relatively short, and surface temperatures would not have increased significantly at this time. As well, the ice advance, if it occurred, would not have altered the permafrost thickness significantly, due to its short duration. In terms of thermal modelling it follows therefore that an ice advance of this type should not appear to have any effect on the permafrost regime - in particular, degradation at the base of the permafrost.

The sea level curve proposed by Hill and Blasco (1986) will be incorporated into the thermal models discussed in the following two chapters. It should be noted that this is the only RSL curve for the Canadian Beaufort Shelf known by the author, and that information is still lacking to construct a sea level curve to account for fluctuations during the past 100,000 years. Carter et al. (1986) suggest that six and possibly seven or eight ~~late~~ Cenozoic transgressions of the Alaskan Arctic Coastal Plain have occurred.

The youngest of these is thought to have occurred between 70,000 and 80,000 years ago. In view of these results, it is important to note that sea level may have fluctuated greatly in the Canadian Beaufort during the Late Pleistocene but sufficient details are not available to confirm or negate this.

5.8 Other Surface Temperature-Related Effects

During the past few thousand years, glacio-eustatic fluctuations in sea level have become minimal and shoreline retreat occurring at a rate of $1-3\text{ m yr}^{-1}$ (Mackay, 1963) has become the dominant land-submerging agent. As a consequence of this recent process, relic permafrost should persist in some nearshore areas irrespective of whether seabottom temperatures are positive or negative. In fact, in the offshore region immediately to the north of Richards Island, permafrost is just as thick as it is onshore, despite warmer seabottom temperatures.

In the Mackenzie Bay region, two additional factors must be considered. High resolution seismic survey data discussed by Blasco (1987) indicate the presence of at least 100m of delta-complex sediments. As these sediments were probably transported by river water with a positive mean annual temperature, any pre-existing permafrost would have been buried by unfrozen sediments. The preservation of relic permafrost in this region, therefore, would depend upon its original thickness, the sedimentation rate and the water temperatures since burial. Knowing that the Iperk Sequence sediments are relatively thin beneath this offshore area, it is unlikely that the original permafrost thickness was great.

Surface temperatures in the emergent modern Mackenzie Delta are significantly higher (-0.4°C to -2.3°C) than those in the old

eastern delta and Tuktoyaktuk Peninsula (-5°C to -10°C) (Hunter et al., 1976). These higher temperatures are a direct consequence of the warm Mackenzie River outflow, and its effect is to restrict permafrost aggradation in submerged areas of this region.

5.9 Discussion

The surface temperature history of the Mackenzie Delta region is clearly very complex. Each of the above mentioned elements which have in the past or are at present acting to modify the ground surface temperature will be taken into account when modelling the permafrost regime.

The following chapter provides the mathematical equations which will be used in the thermal modelling. The application of various past surface temperature histories to the permafrost regime - based on information contained in these preceeding pages - will be carried out in order to reconstruct a paleoclimate which will be capable of accounting for the observed permafrost regime.

FIGURE 11 $\delta^{18}\text{O}$ Variations with Time in Glacier Ice Cores from Camp Century, Greenland (after Dansgaard et al., 1971).

FIGURE 12 Correlation of Paleoclimatic History, Beaufort Sea Continental Shelf, Alaska, and World Sea Level Curve (after Smith, 1986).

FIGURE 13 Location of Quaternary Phenomena, Arctic Coastal Plain, Canada (after Rampton, 1986).

FIGURE 14 Relative Sea Level Curve for the Beaufort Sea Continental Shelf, Canada (after Hill and Blasco, 1986).

CHAPTER 6-THERMAL MODELLING

A thermal model is a mathematical representation of a body which shows how it responds thermally to temperature disturbances. In the present case, the body is the subsurface thermal regime of the Mackenzie Delta which is being acted upon by temperature disturbances at its surface. As was discussed in Chapter 4, heat conduction is the main mode of heat transfer in the subsurface permafrost regime, hence, solutions to Fourier's conduction equation will be sought. Carslaw and Jaeger (1959) derive the mathematical expressions which describe the behaviour of solids to heat conduction. These expressions, therefore, will be referred to in the following sections.

In order to reconstruct a paleoclimate which will be sufficient to account for the observed permafrost regime of the Mackenzie Delta, a time-dependent thermal model is required. This model should incorporate both the lithologic character of the subsurface and the various known surface temperature disturbances which are thought to have acted in the region in the past. By superimposing possible surface temperature histories on the subsurface and monitoring the response of the permafrost to these disturbances, a paleoclimate may be determined to account for the observed thermal regime. It follows, therefore, that the modelled paleoclimate will not be unique. However, it will be sufficient to account for the observed permafrost regime.

6.1 Linear Flow of Heat - Analytical Solution

In order to describe the behaviour of the subsurface to heat conduction through it, a number of assumptions must be made. First,

the flow of heat must be linear. Namely, the lines of flow must be parallel to the Z-axis (where Z is the depth); second, the subsurface is represented by a semi-infinite solid which is a solid bounded by the plane Z=0 (ground surface) and extends to infinity in the positive Z direction; and third, the thermal properties of the subsurface are assumed to be independent of position and temperature. As will be discussed later, this third assumption breaks down when modelling the aggradation and degradation of permafrost.

The equation for the linear flow of heat in one dimension is defined by:

$$\frac{\partial^2 T}{\partial Z^2} - \frac{1}{\alpha} \frac{\partial T}{\partial t} = 0 \quad (28)$$

Equation (28) is satisfied by (Carslaw and Jaeger, 1959):

$$\int_0^Z t^{-1/2} \exp(-z^2/4\alpha t) dz \quad (29)$$

or

$$2\alpha^{1/2} \int_0^{Z/2\sqrt{\alpha t}} \exp(-\xi^2) d\xi^2 \quad (30)$$

Since,

$$\operatorname{erf} x = \frac{2}{\sqrt{\pi}} \int_0^x \exp(-\xi^2) d\xi \quad (31)$$

where "erf x" is the error function of x, then the solution to (28) may be written as:

$$A \operatorname{erf} \frac{x}{2\sqrt{\alpha t}} \quad (32)$$

where A is an arbitrary constant.

The error function has the properties $\operatorname{erf} 0 = 0$ and $\operatorname{erf} \infty = 1$.

If the region $Z > 0$ of the semi-infinite solid is initially at a

temperature $V + kZ$ where V represents a step change in surface temperature and K is a constant, and the plane $Z=0$ is kept at zero temperature, then the temperature T at depth Z and time t is given by (Carslaw and Jaeger, 1959, [2.4] equation (13)):

$$T = V \operatorname{erf} \frac{Z}{2\sqrt{at}} + kZ \quad (33)$$

If the surface temperature is other than zero, Equation (33) is modified to:

$$T = T_0 + V \operatorname{erf} \frac{Z}{2\sqrt{at}} + gZ \quad (34)$$

where g is the geothermal gradient.

If the surface temperature disturbance lasts from times t_2 to t_1 , the temperature is represented by:

$$T = T_0 + V \left(\operatorname{erf} \frac{Z}{2\sqrt{at_1}} - \operatorname{erf} \frac{Z}{2\sqrt{at_2}} \right) + gZ \quad (35)$$

If it is assumed that t_1 is infinity - which would imply that the temperature disturbance acted indefinitely - then Equation (35) would reduce to:

$$T = T_0 + V \left(1 - \operatorname{erf} \frac{Z}{2\sqrt{at}} \right) + gZ \quad (36)$$

and the temperature T could be calculated after a time t .

Using Equation (34), a surface temperature disturbance described by V and acting for a time t , may be imposed on the subsurface thermal regime; and the temperature T may be calculated at different depths Z . In this way, a thermal plot may be produced which will describe the transient response of the subsurface to the step temperature disturbance at the surface. By repeating the process and changing the initial conditions to incorporate the previous results, various step surface temperature changes may be modelled. The final calculated thermal profile will reflect the

entire surface temperature history imposed on the subsurface.

The software (HPBASIC 5.0) listed in Appendix II, was written for use with an HP200/300 series computer, and utilizes Equation (34) in order to model the effect of past climatic variations on the subsurface. The solution is appropriate to step changes in surface temperature for a semi-infinite region $Z > 0$. The temperature at depth Z - in increments of 10m from 0 to 1000m - is calculated in response to a single surface temperature disturbance of magnitude V_1 . The transient response of the subsurface is shown graphically at various times, $t=100$ to 1000 years in increments of 100 years and for $t=10,000$ to 100,000 years in increments of 10,000 years.

This program will be used to show the transient response of the offshore region to submergence. Results for the Beaufort Sea region are discussed in the following chapter.

The major problem associated with this analytical solution is that it does not accurately describe the response of the subsurface where freezing and thawing occur at the base of the permafrost. The reason for this requires a discussion of latent heat.

6.2 Freeze-Thaw Boundaries

When a temperature disturbance is applied to the surface of the earth, the ground responds in a way described in the previous section. If the surface temperature drops, the temperature of the ground will drop gradually until an equilibrium is established. This process requires that heat be removed from the earth - owing to the fact that heat flows from the warmer to the colder areas. It is this loss of heat which results in a drop in temperature of the ground and causes a shift in the thermal gradient.

In a permafrost regime, a slightly more complicated process takes place. A temperature disturbance at the surface will ultimately propagate (although attenuated) down to the base of the permafrost. The base of the permafrost is referred to as a freeze-thaw boundary. At this boundary, phase changes occur in response to even the slightest temperature disturbance. In order for such a phase change to take place, a certain amount of heat must be added to or removed from the ground. This heat is referred to as the Latent Heat of Fusion and has a value of $3.33 \times 10^8 \text{ J kg}^{-1}$.

During a phase change, no change in temperature takes place until such time as the latent heat is removed (during freezing) or added (during thawing). Once the phase change has occurred, the temperature will continue to rise or fall in response to the temperature disturbance.

In a permafrost environment, surface temperature disturbances result in the delayed migration of the permafrost boundary. This requires special consideration when thermal modelling is carried out, and the equations presented in the previous section no longer apply if significant moisture is present. Various approaches exist for determining an analytical solution to the freeze-thaw problem - for example, Sharbatyn (1973) - however, despite attempts made by the author to determine such a solution, none was found.

Both Neumann's solution and Stefan's solution (Ingersoll et al., 1954, chapter 10) for freezing and thawing are appropriate for bodies which are initially at a constant temperature V , and are not easily modified to incorporate bodies which have an initial linear temperature distribution. As a result, a numerical solution to the present problem had to be sought.

6.3 The Permafrost Regime - A Numerical Solution

In order to construct a numerical model of the permafrost regime which incorporates the latent heat involved in the transformation from water to ice (or equivalently from ice to water), it is necessary to look at conditions at the freeze-thaw boundary. Above the boundary, interstitial moisture is predominantly in the form of ice, and below, the moisture is probably in the form of liquid water. Lachenbruch et al. (1982) indicate that the thermal balance at the base of the frozen layer which, in general, will be moving with some velocity v , is given by:

$$G_u K_u - G_f K_f = L N v \quad (37)$$

where G represents the thermal gradient, K represents thermal conductivity, and the subscripts "u" and "f" refer to material below the boundary (unfrozen) and above the boundary (frozen), respectively. L represents the latent heat of fusion, and N represents the porosity (which is assumed to be the same as the ice content).

If the permafrost is degrading, at a rate v , the right side of Equation (37) describes the rate of absorption of heat at the interface between the ice and water. This must be balanced by the heat conducted upward toward, and upward away from, the interface (left side). Of course, if the permafrost is neither aggrading or degrading, then v is approximately zero and the heat flow is continuous across the interface.

Lachenbruch et al. (1982) derive the expressions used to estimate the rate of upward thinning of permafrost. It is assumed that the region below remains in a quasi steady-state during the process and, consequently, the heat supplied to the base of the

permafrost remains constant and equal to the geothermal flux Q . Therefore, it follows from Equation (37) that the heat balance at the base of the frozen layer requires:

$$-L N \frac{dZ}{dt} = Q - K_f \frac{dT}{dZ} \quad (38)$$

where dZ/dt represents the velocity, Q replaces the product of the thermal gradient and the thermal conductivity of the thawed layer (as this will remain constant during the thawing process), and dT/dZ represents the thermal gradient in the frozen layer.

Although Equation (38) was used by Lachenbruch et al. (1982) for thinning of permafrost in the offshore, it applies mathematically to freeze-thaw boundaries and, therefore, adequately approximates the rate of downward aggradation of permafrost.

In order to determine the temperature distribution in the ground which results from a surface temperature disturbance, one might consider finding the temperature T in a layer of material of constant thickness Z with a basal temperature T_z if it is initially in equilibrium with a surface temperature T_0 and that surface temperature is suddenly raised to T_s at time $t=0$.

The solution is the sum of a transient part T_1 and a steady-state part T_2 such that:

$$T(z,t) = T_1(z,t) + T_2(z) \quad 0 < z < Z \quad (39)$$

where

$$T_2 = T_s + (T_z - T_s)z / Z \quad (40)$$

and T_1 is the solution to Fourier's conduction equation in one dimension subject to the conditions:

$$T_1(z,0) = (T_0 - T_s)(1 - z/Z) \quad t=0 \quad (41)$$

$$T_1(z,t) = 0 \quad z=0 \text{ and } Z \quad t>0 \quad (42)$$

The solution of the transient part is given by Lachenbruch et al. (1982) and is represented by:

$$T_1(z, t) = (T_0 - T_s) \frac{2}{\pi} \sum_{n=1}^{\infty} \frac{1}{n} \exp\left(-\frac{n^2 \pi^2 t}{4\lambda}\right) \sin\left(n\pi \frac{z}{Z}\right) \quad (43)$$

Equation (38) may therefore be written to incorporate (39) and (40) above such that:

$$-L N \frac{dZ}{dt} = Q - K_f \frac{dT_1}{dZ} - K_f \frac{T_z - T_s}{Z(t)} \quad (44)$$

For values of $t > \lambda$ where $\lambda = Z^2/4\alpha$ and α is the thermal diffusivity of the frozen material, the derivative on the right in Equation (44) is small and may be excluded. As the base of the permafrost varies through the process, so will the value of the time constant (λ). Therefore, in order to comply with the above conditions, the time interval t - over which the process occurs - must vary accordingly.

An assumption is therefore made that the transient behaviour - described by the excluded derivative in Equation (44) - of the ground to surface temperature disturbances, results in negligible permafrost variations for values of $t > \lambda$ (refer to discussion by Lachenbruch et al., 1982).

Equation (44) has been used to model the effect of surface temperature disturbances on the permafrost regime. The temperature at the base of the permafrost is taken to be 0°C although freezing and thawing may occur at temperatures other than this value (refer to Chapter 2).

The computer program listed in Appendix III uses Equation (44) to calculate the change in the depth of permafrost (dZ) as a result of a step increase or decrease in surface temperature acting over a time interval (dt). The initial steady-state depth of the permafrost is first calculated using the assumed regional heat flow, the

conductivity of the subsurface material and the initial surface temperature. Using this depth, the value of the time constant is calculated.

A step change in surface temperature is then applied to the region and the time interval over which it acts is specified. The change in the base of the permafrost is calculated using a time $t = \lambda$ in Equation (44), and then the new permafrost base is calculated. This numerical iteration continues - recalculating the time constant with each change in the base of the permafrost - until such time as the sum of the individual time constants is equal to or greater than the initial specified time interval.

This process is repeated until the entire surface temperature history has been applied to the region, at which point a final calculated permafrost thickness may be compared to the observed average permafrost thickness for the region.

The model is suited to problems where the initial permafrost thickness is known, although it may be applied, as it was here, to reconstruct a possible paleoclimate from estimated initial conditions. As well, the model is restricted to use in problems where the transient response of the subsurface to surface temperature disturbances is not required.

The application of this numerical model to the Mackenzie Delta in order to reconstruct a paleoclimate has been carried out for the five sub-regions identified in Chapter 4. The modelling results are discussed in the following chapter.

CHAPTER 7 - DISCUSSION

The reconstruction of a paleoclimate which will be sufficient to account for the observed distribution and thickness of permafrost in the Mackenzie Delta (Figure 3) is a process which must be carried out by trial and error. Various surface temperature disturbances which act for specified time intervals may be applied to the region and the response of the subsurface in terms of the variation in the thickness of permafrost may be determined. The primary objective of the modelling is to incorporate all suspected fluctuations in surface temperature - discussed in Chapter 5 - to obtain a final calculated permafrost depth which is comparable to the observed average permafrost thicknesses in the sub-regions.

The initial surface conditions must be established, therefore a major assumption must be made. This requires that any Illinoian permafrost must have thawed completely during the Sangamonian Interglacial. As a result, the paleoclimate - beginning in the Late Pleistocene and lasting to the Holocene - must be sufficient to account for the permafrost regime. This assumption is a likely one, owing to the fact that during the Sangamonian, surface temperatures were probably high enough and acted for sufficient time to thaw any permafrost which may have been present. Given an initial permafrost thickness of 800m at the end of the Illinoian - a thickness similar to that present at the end of the Wisconsinan - then, if Sangamonian surface temperatures were -1°C for approximately 50,000 years, less than 250m of permafrost would remain at the end of the Sangamonian. If surface temperatures were at any time warmer than -1°C for a period of more than 2,000 years, it is likely that no permafrost

would remain.

For modelling purposes the surface temperature history is best separated into two components; namely, a component which involves the determination of the time intervals of the paleoclimatic history, and a component which determines the temperatures associated with each time interval. In order to simplify the modelling process, the time intervals proposed by Smith (1986) were tested. Superimposed on these intervals are surface temperatures for the onshore areas which were based on 1) the oxygen isotope profile for the Greenland Ice Cap (Dansgaard et al., 1971) and 2) amino acid surface temperature estimates for the Beaufort Sea continental shelf (Brigham and Miller, 1983; Smith, 1986). These surface temperatures reflect Late Pleistocene and Holocene paleo-temperatures which may have characterized the Mackenzie Delta region.

For the modern Mackenzie Delta and the offshore Beaufort shelf regions, additional surface temperature information reflecting the warming associated with the Late Pleistocene Glaciation and submergence, respectively, is required.

Again, it should be noted that the modelling temperatures represent mean annual ground surface temperatures which may be of the order of 2°C warmer than the mean annual air temperatures in the region.

Modelling results for the five sub-regions - identified in Chapter 4 - are discussed in the following sections. Values of thermal conductivity, porosity, heat flow and diffusivity are summarized in Table 7 (Chapter 4).

7.1 Onshore Permafrost Regime

The three onshore regions which were modelled include: 1) the Kugmallit Bay region, 2) the Tuktoyaktuk Peninsula and 3) the western portion of Richards Island. Each of these sub-regions reflects a different subsurface lithologic regime; however the paleoclimate is believed to have been similar for each.

These onshore regions are assumed to have remained emergent from the Beaufort Sea during the Late Pleistocene. In addition, the regions would have been covered by an ice sheet during the Early Wisconsinan (Buckland Glaciation, Rampton, 1982) which is estimated to have had a duration of approximately 5,000 years. The surface temperature prior to and following the ice advance - during the Early and Late Wisconsinan - would necessarily have been extremely cold, and the model indicates that a surface temperature of -18°C during these intervals would be sufficient to account for the observed permafrost thicknesses in the sub-regions. It should be pointed out that the model is sensitive to temperature fluctuations of a degree or more because the time intervals are so long. If Early and Late Wisconsinan surface temperatures were much colder than -18°C , the model indicates that too much permafrost would have aggraded and this permafrost cannot be accounted for. Likewise, surface temperatures warmer than -18°C result in too little permafrost.

This estimate agrees with that proposed by Brigham and Miller (1983) and with Smith (1986) for the Alaskan Beaufort Shelf. A Mid-Wisconsinan rise in temperature to approximately -8°C , similar to mean surface temperature today, is similarly proposed. Holocene

temperatures are assumed to have remained constant at -10°C during the past 15,000 years.

Figure 15 displays graphically the modelling results for the onshore regions. The first graph illustrates the surface temperature history outlined above which was applied to each of the sub-regions below. The variations in the base of the permafrost (0°C isotherm) as a result of the successive temperature disturbances, are shown graphically for each sub-region.

In the Kugmallit Bay region, a representative depth to the base of the ice-bearing permafrost is estimated at approximately 700m (refer to Figure 3), although values slightly higher than this also characterize the region. The permafrost thickness (depth to the 0°C isotherm) calculated using the described paleoclimate is slightly less than 800m. The discrepancy which exists between these two estimates may be a result of a slight over-estimation of thermal conductivity used in the model. The value of $4.5 \text{ W m}^{-1} \text{ K}^{-1}$ (refer to Table 7) which was calculated from average gradients in the area has a standard deviation of $1.6 \text{ W m}^{-1} \text{ K}^{-1}$; hence, this value may not be representative of the subsurface thermal regime. It should be pointed out that iterations in the thermal models may contribute to slight errors in the final results for all regions.

In the Tuktoyaktuk Peninsula region, a representative depth to the base of the ice-bearing permafrost is estimated to be in the range of 500 to 600m. Modelling results indicate a permafrost thickness of slightly greater than 700m. One possible reason for this discrepancy may be the result of glaciation. As discussed in Chapter 5, the extent and duration of ice cover during the Late Pleistocene remains debatable.

With reference to the glaciation, two factors may contribute to an increased surface temperature which may not have been incorporated into the thermal model. First, it is possible that the Buckland ice sheet covered the area for a period longer than 5000 years as proposed in the model. Second, if during the Mid- to Late Wisconsinan an ice sheet was present along the Tuktoyaktuk Peninsula (Sabine Phase, Rampton, 1982) and remained there for a sufficient length of time, surface warming would have resulted in degradation of permafrost - particularly if the ice sheet remained in the region for a period greater than 2000 years.

These two factors, if considered separately or together may account for the discrepancy between the modelling results and the observed permafrost thicknesses. Incorporating these two factors was not carried out due to the uncertainty of the duration and ice cover of the ice sheets.

In the western portion of Richards Island, a representative depth to the base of the ice-bearing permafrost is estimated at approximately 350m. The thermal model indicates a permafrost thickness of slightly less than 400m. This estimate agrees with the observed permafrost thicknesses in the region, particularly if the difference between the IBPF base and the 0°C isotherm depth is considered. In a mudstone such as the Richards Sequence, this discrepancy becomes more pronounced due to the higher unfrozen water content.

In all three cases, the final calculated permafrost thickness compares reasonably well with the observed IBPF bases in the regions. The proposed paleoclimate for these three sub-regions is sufficient to account for the observed average permafrost

thicknesses in the regions.

7.2 The Modern Mackenzie Delta

The modern Mackenzie Delta was considered separately due to the uniqueness of its recent surface temperature history. From the Late Sangamonian to the end of the Mid-Wisconsinan, the surface temperature was probably similar to that in the onshore regions discussed above. The exception involves the Late Wisconsinan - in particular, the length of time of emergence following deglaciation. The limit of Late Wisconsinan ice proposed by Rampton (1986) extends slightly into the modern Mackenzie Delta, therefore, during deglaciation, which would have begun somewhere around 20,000 years B.P., glacial meltwater would have covered the area. This glacial meltwater has been assigned a temperature of $+1^{\circ}\text{C}$.

In addition, a Holocene surface temperature of $+1^{\circ}\text{C}$, representative of Shallow Bay in the modern Delta, has been applied to the region. This positive value of course applies to areas which are submerged beneath warm waters of the Mackenzie River such that degradation of permafrost occurs. Despite this, those areas which are at present emergent have aggrading permafrost and temperatures of -0.4°C to -2.3°C and thin permafrost at present.

The surface temperatures associated with the modern Delta are constantly changing due to the migration of river channels and the variation of sediment areas of deposition. This region, therefore, is thermally unstable and highly variable and it is difficult to determine a typical model accurately.

The surface temperature history applied to the region and the

resulting permafrost variations are shown graphically in Figure 16. As expected - due to the relatively warm Holocene temperatures - the permafrost thickness is significantly less than that modelled in the Tuktoyaktuk Peninsula area which has a similar subsurface lithologic regime.

In the modern Mackenzie Delta, a representative depth to the base of the ice-bearing permafrost is estimated at less than 100m, although permafrost may be completely absent in some areas. The calculated permafrost thickness in the region is approximately 200m. This value, despite being slightly greater than the observed average value, is a relatively good estimate. The paleoclimate which has been applied to the sub-region, therefore, is sufficient to account for the observed permafrost regime.

7.3 The Offshore Beaufort Shelf

To illustrate the effect of submergence on the permafrost terrain, a simple model is presented for the offshore region. Using the computer program listed in Appendix II, the effect of submergence to a permafrost regime is modelled. Initially, the ground surface is at a temperature of -9°C - similar to temperatures on the northern portion of Richards Island - the thermal gradient is $13.7^{\circ}\text{C km}^{-1}$ - calculated using the steady state relationship between a thermal conductivity of $4.5 \text{ W m}^{-1}\text{K}^{-1}$ and a heat flow of 60 mW m^{-2} - and the thermal diffusivity is $0.017 \text{ cm}^2 \text{ s}^{-1}$. At time T_0 the area is inundated by the sea with a mean annual bottom water temperature of -1.8°C .

Figure 17 shows the effect of the surface temperature

disturbance plotted for various times down to a depth of 1000m. This graph clearly indicates the relatively small amount of degradation at the base of the permafrost from the top of the permafrost arising in the short term due to submergence. Only after many thousands of years does the permafrost thaw - this being a consequence of basal melting from the geothermal heat flow from the earth's interior. Lachenbruch et al. (1982) state:

"Therefore, submergence is followed by a period in which the 'triangular cold reserve' is depleted by rapid conduction of heat downward from the warm seabed at $z=0$ and the continued influx of heat from below (at $z=Z$)."

It should be noted that this program does not incorporate the effect of latent heat.

The paleoclimate which has been reconstructed to account for the permafrost regime in the offshore - in particular the offshore region to the north of Richards Island - is shown graphically in Figure 18. The offshore seabottom has been given thermal parameters similar to those characterizing the northeastern portion of Richards Island with the exception that water depths are approximately 60m. These depths imply that the modelled offshore region has been submerged for the past 10,000 years (refer to Figure 14, after Hill and Blasco, 1986).

Following the Late Wisconsinan, the offshore was exposed to Holocene surface temperatures similar to those experienced by other portions of the Mackenzie Delta. This exposure lasted approximately 5,000 years after which time submergence would have begun.

Apart from the rise in surface temperature associated with submergence during the past 10,000 years, two additional periods of

submergence - lasting throughout the Mid-Wisconsinan and the Sangamonian - have been incorporated. Other than these, the paleoclimate is similar to that applied to the onshore regions.

At present a representative depth to the base of the ice-bearing permafrost at a depth of 60m below sea level, is approximately 500m. This value, in general, coincides with the calculated thickness obtained from the thermal model (Figure 18).

7.4 Suggestions for Future Research

The study which was undertaken in the Mackenzie Delta in order to reconstruct a paleoclimate sufficient to account for the observed permafrost occurrence was a large-scale one. Regional variations in the permafrost thickness and distribution were described in terms of the subsurface thermal regime and the surface temperature history which affected the region at large.

A comprehensive study of more localized permafrost variations is the next logical step to describing the permafrost regime in the Mackenzie Delta. In terms of lithologic variations, a more detailed study of the thermal properties should be undertaken. In particular, accurate measurements of thermal conductivity could be made, either by laboratory investigations where core material where core material exists. This would help to limit the range of thermal conductivity exhibited within a sequence such that more representative values could be determined. In addition, accurate thermal conductivities could be incorporated into thermal models applied to smaller regions.

In terms of glaciation, a better understanding of the glacial

extent and duration is required. As was stated earlier, there still exists some doubt as to the possibility of a late Early Wisconsinan or Mid-Wisconsinan glaciation (Sabine Phase). If a glaciation did actually occur in the region, surface temperatures may have been significantly different from those presented here in the paleoclimatic model. As well, the author questions the age of the Buckland limit proposed by Rampton (1986) which is estimated at Early Wisconsinan. It is believed that this age was based on insufficient data and that the Mason River limit (Rampton, 1982) on the Tuktoyaktuk Peninsula, may in fact be Illinoisan in age. This could ultimately alter the age assignment of Early Wisconsinan to the Buckland limit elsewhere in the region. Michel (1982) obtained an age of approximately 13,000 years for organics located within till-like diamicton deposits and approximately 24,000 years for organics within underlying sands in the Illisarvik area on the northeastern coast of Richards Island. This could imply a Late Wisconsinan ice cover in the area, and that the Late Wisconsinan limit occurred further to the north than is expected at present. Although the observed permafrost thicknesses indicate that Late Wisconsinan ice is unlikely to have advanced and remained for any length of time in the area, a short-lived ice advance may, in fact, have been present.

As was stated in Chapter 5, Hill et al. (1985) indicate that a Mid-Wisconsinan ice sheet is suspected to have covered the Mackenzie Bay region - this being inferred from the relative sea level curve for the Canadian Beaufort Shelf. It is suggested that the Early Wisconsinan ice may not have receded. This, of course, would imply a surface temperature history much warmer than that proposed here.

Offshore, a more extensive investigation of the effect of salt migration on the degradation of permafrost could be undertaken. Harrison and Osterkamp (1981) suggest that this may be a major component in the degradation of offshore permafrost. If this is the case, then modeling should not be based on thermal aspects alone.

In terms of thermal modelling, two additional variations are suggested. The first requires a more detailed mathematical treatment of the permafrost regime in the sense that a multi-layered subsurface could be modelled. As such, variations with depth of the thermal properties could be incorporated. This method would greatly improve upon the model as the boundaries between the various sequences could be examined. This problem proved too difficult to solve for a permafrost regime in the present instance as finite element numerical models are called for.

Finally, longer- and shorter-term paleoclimates could be modelled. For a long-term study, the aggradation of permafrost during the Illinoian and the subsequent degradation of permafrost during the Sangamonian could be modelled. This would provide better defined initial conditions with which to model the Wisconsinan variations.

In the short term, an investigation of the paleoclimate during the Late Wisconsinan and Holocene may be undertaken in more localized areas such as the Parsons Lake area in the southern portion of the delta. Part of this region was glaciated during the Late Wisconsinan and part remained unglaciated. A detailed study of permafrost variations would provide an interesting supplement to the information presented in this thesis.

Clearly, there are many possibilities for future study. This-

thesis forms a starting point for more detailed research in the Mackenzie Delta region and presents a number of interesting points concerning the paleoclimate and associated permafrost regime.

FIGURE 15 Surface Temperature History (a) Applied to Onshore Regions and Permafrost Variations in (b) Kugmallit Bay, (c) Tuktoyaktuk Peninsula and (d) Western Richards Island regions.

FIGURE 16 Surface Temperature History (a) and Resulting Permafrost Variations (b) in the Modern Mackenzie Delta.

FIGURE 17 Transient Response of Offshore Region to Submergence for Various Times t following Submergence.

FIGURE 18 Surface Temperature History (a) and Resulting Permafrost Variations (b) in the Offshore Beaufort Shelf

SUMMARY

In summary, a number of conclusions may be drawn from the analysis of the permafrost regime in the Mackenzie Delta. These include: 1) the depth to the base of the ice-bearing permafrost varies significantly throughout the region, 2) there exists a strong correlation between the Pliocene-Pleistocene Iperk sequence and the occurrence of deep permafrost, 3) Early and Late Wisconsinan surface temperatures of -18°C are required to account for the observed thicknesses of permafrost in the region, 4) a Mid-Wisconsinan rise in temperature to -8°C is proposed, 5) an Early Wisconsinan ice sheet covered the region for at least 5000 years and finally, 6) present permafrost conditions indicate that a Late Wisconsinan ice advance is unlikely to have been extensive.

REFERENCES

- ALLEN, D., MICHEL, F. and JUDGE, A. (1988) Paleoclimate and permafrost in the Mackenzie Delta (in press). In, Proceedings of the Fifth International Conference on Permafrost, Trondheim, Norway.
- APTIKAEV, F.F. (1964) Temperature field effect on the distribution of seismic velocities in the permafrost zone. In, Akad. Nauk SSSR Sibjrskoe otd-ie Inst. merzlotoredeniia. Teplovye protsessy V merzlykh porod. (In Russian).
- BADA, J.L. (1972) The dating of fossil bones using the racemization of isoleucine, Earth and Planetary Science Letters, 15, p. 223-231.
- BIRCH, F. (1950) Flow of heat in the Front Ranges, Colorado. Bulletin of Geological Society of America, 61, p. 567-630.
- BLASCO, S.M. (1987) Wisconsinan ice advance in the Mackenzie Trough of the Canadian Beaufort Sea (abstract). Presented at the XII International Congress International Union for Quaternary Research, Ottawa, Canada, July, 1987.
- BRIGHAM, J.K. (1982) Comment on comparison of uranium series, radiocarbon and amino acid data from marine molluscs, Baffin Island, Arctic Canada. Geology, 10, p. 215.
- BRIGHAM, J.K. (1983) Stratigraphy, amino acid geochronology, and correlation of Quaternary sea-level and glacial events, Broughton Island, Arctic Canada. Canadian Journal of Earth Sciences, 20, p. 577-598.
- BRIGHAM, J.K. and MILLER, G.H. (1983) Paleotemperature estimates of the Alaskan Arctic Coastal Plain during the last 125,000 years. In Proceedings of the Fourth International Conference on Permafrost, Fairbanks, Alaska, p. 80-85.
- BURGESS, M. (1977) Shallow thermal studies, Shallow Bay, Mackenzie Delta, Internal Report No. 80-10, Earth Physics Branch, E.M.R., 27 pp.
- CARSLAW, H.S. and JAEGER, J.C. (1959) Conduction of Heat in Solids. Oxford University Press, New York, 386 pp.
- CERMAK, V. (1971) Underground temperature and inferred climatic temperature of the past millenium. Paleogeography, Paleoclimatology, Paleoecology, 10, p. 1-19.
- D & S CONSULTING GROUP LTD. (1983) Study of well logs in the Mackenzie Delta-Beaufort Sea area to outline permafrost thickness and/or gas hydrate occurrence. Earth Physics Branch Open File Number 83-10, 242 pp.

DANSGAARD, W., JOHNSEN, S.J., CLAUSEN, H.B. and LANGWAY, C.C.Jr. (1971) Climatic record revealed by the Camp Century core ice. In *The Late Cenozoic Glacial Ages*, ed. Turkian, Yale University Press, p. 37-56.

DIETRICH, J.R., DIXON, J. and MCNEIL, D.H. (1985) Sequence analysis and nomenclature of Upper Cretaceous to Holocene strata in the Beaufort-Mackenzie Basin. In *Current Research, Part A*, Geological Survey of Canada, Paper 85-1A, p. 613-628.

DIXON, J. (1982) Jurassic and Lower Cretaceous subsurface stratigraphy of the Mackenzie Delta-Tuktoyaktuk Peninsula, N.W.T., Geological Survey of Canada Bulletin 349, 52 pp.

DIXON, J., MCNEIL, D.H., DIETRICH, J.R., BUJAK, J.P. and DAVIES, E.H. (1984) Geology and biostratigraphy of the Dome Gulf et al. Hunt Kopanoar M-13 Well, Beaufort Sea. Geological Survey of Canada, Paper 82-13, 28 pp.

DIXON, J., DIETRICH, J.R., MCNEIL, D.H., MCINTYRE, D.J., SNOWDON, L.R. and BROOKS, P. (1985) Geology, biostratigraphy and organic geochemistry of Jurassic to Pleistocene strata, Beaufort-Mackenzie area, Northwest Canada. Canadian Society of Petroleum Geologists, Course Notes.

DIXON, J. (1986) Cretaceous to Pleistocene Stratigraphy and Paleogeography, Northern Yukon and Northwestern District of Mackenzie. Bulletin of Canadian Petroleum Geology, 34.1, p. 49-70.

GOLD, L.W. and LACHENBRUCH, A.H. (1973) Thermal conditions in permafrost - A review of North American literature. In, *Proceedings of the Second International Conference on Permafrost*, Yakutsk, July 1973, North American Contribution, Washington, D.C. p. 3-25.

HARRISON, W.D. and OSTERKAMP, T.E. (1981) Theoretical models for sub-sea permafrost. Geophysical Institute of Alaska, Fairbanks, Alaska.

HATLELID, W.G. and MACDONALD, J.R. (1979) Permafrost determination by seismic velocity analysis. In *Proceedings of Symposium on Permafrost Field Methods and Permafrost Geophysics*, Assoc. Comm. Geotechnical Research, Technical Memo 124, p. 146-147.

HAWKINGS, T.J. and HATLELID, W.G. (1975) The regional setting of the Taglu gas field. In, Glass, D.J., Yorath, C.J. and Parker, E.R. (Eds.), *Canada's Continental Margins*, Canadian Society of Petroleum Geologists, Memoir 4, p. 633-648.

HILL, P.R., MUDIE, P.J., MORGAN, K. and BLASCO, S.M. (1985) A sea-level curve for the Canadian Beaufort Shelf. Canadian Journal of Earth Sciences, 22, p. 1383-1393.

- HILL, P.R. and BLASCO, S.M. (1986) Implications of a Beaufort Sea relative sea-level curve to Arctic Mid and Late Wisconsinan Glacial Events. In Correlation of Quaternary Deposits and Events Around the Margin of the Beaufort Sea: Contributions from a Joint Canadian-American Workshop, April, 1984, J.A. Heginbottom and J-S. Vincent (Eds.), Geological Survey of Canada, Open File Report 1237, 60 pp.
- HNATIUK, J. and RANDALL, A.G. (1977) Determination of permafrost thickness in wells in northern Canada. Canadian Journal of Earth Sciences, 14, p.375-383.
- HOEKSTRA, P. and MCNEILL, D. (1973) Electromagnetic probing of permafrost. In, Proceedings of the Second International Conference on Permafrost, Yakutsk, July 1973, North American Contribution, Washington, D.C.
- HUNTER, J.A.M. (1973) The application of shallow seismic methods to mapping frozen surficial materials. In, Proceedings of the Second International Conference on Permafrost, Yakutsk, July 1973, North American Contribution, Washington, D.C.
- HUNTER, J.A.M., JUDGE, A.S., MACAULAY, H.A., GOOD, R.L., GAGNE, R.M. and BURNS, R.A. (1976) Permafrost and sub-seabottom materials in the southern Beaufort Sea. Technical Report No. 22, Beaufort Sea Project, Department of the Environment, 174 pp.
- INGERSOLL, L.R., ZOBEL, O.J. and INGERSOLL, A.C. (1954). Heat Conduction with Engineering, Geological and Other Applications. University of Wisconsin Press, Madison, 325 pp.
- JACKSON, R.D. and TAYLOR, S.A. (1965) Heat transfer, Agronom. Monograph 9, Part 1, American Society of Agronomists, p. 349-360.
- JOHANSEN (1972), (1973) Beregningsmetode for varmeledningsevne av fuktige og frosne jordarter. Del I, Teoretisk grunnlag. Frost i Jord., 7, p. 17-25. Del II, Frost i Jord., 10, p. p. 13-28. (Includes English Summaries: A Method for Calculation of Thermal Conductivity of Soils, Part 1, Part 2).
- JOHANSEN (1975) Varmeledningsevne av jordarter (Thermal Conductivity of Soils). Inst. f. Kjøleteknikk, Trondheim, 231 pp. (with English summary).
- JONES, P.B., BRACHE, J. and LENTIN, J.K. (1980) The geology of the 1977 offshore hydrocarbon discoveries in the Beaufort-Mackenzie Basin, N.W.T. Bulletin of Canadian Petroleum Geology, 28, 1, p. 81-102.
- JUDGE, A.S. (1973) The thermal regime of the Mackenzie Valley: Observations of the Natural State. Government of Canada, Internal Report to the Environment - Social Committee on northern pipelines, 170 pp.

JUDGE, A.S. (1974) The occurrence of offshore permafrost in northern Canada. In, The Coast and Shelf of the Beaufort Sea, Washington, D.C. Arctic Institute of North America, p. 427-437

JUDGE, A.S., TAYLOR, A.E. and BURGESS, M. (1979) Canadian Geothermal Data Collection - Northern Wells, 1977-78. Geothermal Series Number 11, Earth Physics Branch, E.M.R., 188 pp.

JUDGE, A.S., TAYLOR, A.E., BURGESS, M. and ALLEN, V.S. (1981) Canadian Geothermal Data Collection - Northern Wells, 1978-80. Geothermal Series Number 12, Earth Physics Branch, E.M.R., 190 pp.

JUDGE, A.S. and TAYLOR, A.E. (1985) Permafrost distribution in northern Canada: Interpretation of well logs. In, Workshop on Permafrost Geophysics, Golden, Colorado, Special Report 85-5.

JUDGE, A.S. (1986) Permafrost distribution and the Quaternary history of the Mackenzie-Beaufort region: A geothermal perspective. In Correlation of Quaternary Deposits and Events Around the Margin of the Beaufort Sea: Contributions from a Joint Canadian-American Workshop, April, 1984, J.A. Heginbottom and J-S. Vincent (Eds.), Geological Survey of Canada, Open File Report 1237, 60 pp.

KURFURST, P.J. and KING, M.S. (1972) Static and dynamic elastic properties of two sandstones at permafrost temperatures. Transactions of the Society of Petroleum Engineers of the American Institute of Mining, Metallurgical and Petroleum Engineers, 253, p. 495-504.

LAGHENBRUCH, A.H., SASS, J.H., MARSHALL, B.V. and MOSES, T.H. Jr. (1982) Permafrost, heat flow and the geothermal regime at Prudhoe Bay, Alaska. Journal of Geophysical Research, 87, B11 pp. 9301-9316.

LEFLANC-ROSS, D. (1986) An Analysis of the Occurrence of Permafrost in the Mackenzie Delta-Beaufort Sea Region, N.W.T. B. Sc. Thesis, Department of Geology, Carleton University, Ottawa, 78 pp.

MACKAY, J.R. (1963) The Mackenzie Delta area, Northwest Territories, Canada: Department of Mines and Technical Surveys, Geographical Branch, Memoir No. 8, 202 pp.

MACKAY, J.R. (1972) Offshore permafrost and ground ice, southern Beaufort Sea, Canada. Canadian Journal of Earth Sciences, 9, p. 1550-1561.

MACKAY, J.R., RAMPTON, V.N. and FYLES, J.V. (1983) Relic Pleistocene permafrost, western Canadian arctic, Canada. Science, 176, p. 1321-1323.

- MACKAY, J.R. and MATTHEWS, J.V. Jr. (1983) Pleistocene ice and sand wedges, Hooper Island, Northwest Territories. Canadian Journal of Earth Sciences, 20, p. 1087-1097.
- NEAVE, K.G., JUDGE, A.S., HUNTER, J.A. and MACAULAY, H.A. (1978) Offshore permafrost distribution in the Beaufort Sea as determined from temperature and seismic observations. In Current Research, Part C, Geological Survey of Canada, Paper 78-1C, p. 13-18.
- NENTWICH, F.W. and YOLE, R.W. (1982) Sedimentary petrology and stratigraphic analysis of the subsurface Reindeer Formation (Early Tertiary) Mackenzie Delta-Beaufort Sea area, Canada. Arctic Geology and Geophysics, A.F. Embry and H.R. Balkwell (Eds.), Canadian Society of Petroleum Geologists, Memoir 8, p. 55-81.
- OSTERKAMP, T.E. and PAYNE, M.W. (1981) Estimates of permafrost thickness from well logs in northern Alaska. In Cold Regions Science and Technology, 5, p. 13-27.
- PANDIT, B.I. and KING, M.S. (1979) A study of the effects of pore-water salinity on some physical properties of sedimentary rocks at permafrost temperatures. Canadian Journal of Earth Sciences, 16, p. 1566-1580.
- PREST, V.K. (1969) Retreat of Wisconsin and Recent ice in North America. Geological Survey of Canada, Map 1257A, Scale 1:5 000 000.
- RAMPTON, V.N. (1982) Quaternary geology of the Yukon Coastal Plain. Geological Survey of Canada, Bulletin 317, 49 pp.
- RAMPTON, V.N. (1986) Quaternary history of the Arctic Coastal Plain in Canada. In, Correlation of Quaternary Deposits and Events Around the Margin of the Beaufort Sea, Contributions from a Joint Canadian-American Workshop, April 1984, J.A. Heginbottom and J-S. Vincent (Eds.), Geological Survey of Canada, Open File Report 1237, p. 10-11.
- SCHROEDER, R.A. and BADA, J.L. (1976) A review of the geochemical applications of the amino acid reaction. Earth Science Reviews, 12, p. 347-391.
- SHEARER, J.M., MACNAB, R.F., PELLETIER, B.R., and SMITH, T.B. (1971) Submarine pingoes in the Beaufort Sea, Science, 175, p. 816-818.
- SMITH, M.W. and HWANG, C.T. (1973) Thermal disturbance due to channel shifting, Mackenzie Delta, N.W.T. In, Proceedings of the Second International Conference on Permafrost, Yakutsk, July 1973, North American Contribution, Washington, D.C.

- SMITH, P.A. (1986) The Late-Pleistocene - Holocene stratigraphic record, Canning River Delta region, Northern Alaska. In Correlation of Quaternary Deposits and Events Around the Margin of the Beaufort Sea: Contributions from a Joint Canadian-American Workshop, April, 1984, J.A. Heginbottom and J-S. Vincent (Eds.), Geological Survey of Canada, Open File Report 1237, 60 pp.
- TAYLOR, A.E. and JUDGE, A.S. (1977) Canadian Geothermal Data Collection - Northern Wells, 1976-77. Geothermal Series Number 10, Earth Physics Branch E.M.R., 194 pp.
- TAYLOR, A. and JUDGE, A. (1983) Shoreline regression: Its effect on permafrost and the geothermal regime, Canadian Arctic Archipelago. In, Proceedings of the Fourth International Conference on Permafrost, Fairbanks, Alaska, National Academy Press, Washington, D.C. p. 1239-1244.
- TAYLOR, A.E., BURGESS, M., JUDGE, A.S. and ALLEN, V.S. (1982) Canadian Geothermal Data Collection - Northern Wells, 1981. Geothermal Series Number 13, Earth Physics Branch, E.M.R., 153 pp.
- TIMUR, A. (1968) Velocity of compressional waves in porous media at permafrost temperatures. Geophysics, 33, p. 584-595.
- VILKS, G., WAGNER, F.J.E. and PELLETIER, B.R. (1979) The Holocene marine environment of the Beaufort Shelf. Geological Survey of Canada, Bulletin 303.
- WEAVER, J.S. and STEWART, J.M. (1982) In situ hydrates under the Beaufort Sea shelf. In, Proceedings of Fourth Canadian Permafrost Conference, Calgary, Alberta, March 2-6, 1981. Ed. H.M. French, National Research Council of Canada, Ottawa, Ontario, p. 312-319.
- WALKER, J.D. and STUART, A.J. (1976) Permafrost investigation by crystal cable surveys, Mackenzie Delta, N.W.T. Transactions SPWLA, 17th Annual Logging Symposium.
- WILLIAMS, P.J. (1982) The Surface of the Earth: An Introduction to Geotechnical Science. Longman Inc., New York, 212 pp.
- YOUNG, S.E. and JUDGE, A.S. (1985) A ground temperature data collection for northern Canada. Presented at the McGill Workshop on Climate and Thermal Regimes in Northern Canada, November 8, 1985, 17 pp.
- YOUNG, F.G. and MCNEIL, D.H. (1984) Cenozoic stratigraphy of the Mackenzie Delta, Northwest Territories, Geological Survey of Canada, Bulletin 336, 63 pp.
- YOUNG, F.G., MYHR, D.W. and YORATH, C.J. (1976) Geology of the Beaufort Mackenzie Basin. Geological Survey of Canada, Paper 76-11, 65 pp.

APPENDIX I

LOCATION NUMBER	SITE NAME	COORDINATES	0°C		METHOD
			ISOTHERM	IBPF	
1	Union Aklavik F-17	68°06'20" 135°04'00"		<161m	
2	Union Aklavik F-38	68°07'15" 135°09'11"		<841m	Res
3	Shell Aklavik A-37	68°16'15" 135°07'46"		103m	Res
4	CPOG Crossley K-60	68°29'39" 129°29'14"		<215m	Res
5	Union Wolverine H-34	68°23'19" 130°38'00"		<471m	Res
6	Amoco Inuvik D-54	68°23'13" 133°44'25"		<384m	Res
7	Shell Napoiak F-31	68°20'25" 134°53'49"		97m	Res
8	Shell Beaver House H-13	68°22'16" 135°33'03"	E197	135m	Res
9	CPOG Kugaluk N-02	68°31'55" 131°31'19"			
10	Gulf East Reindeer P-60	68°39'45" 133°43'00"		158m	Res
11	Esso Napartok M-01	68°30'47" 134°32'18"		71m	Res
12	Chevron Fish River B-60	68°39'03" 136°13'39"		<232m	Res
13	Gulf East Reindeer C-38	68°47'10" 133°39'15"		<490m	Res
14	Gulf Ogeoqueeq J-06	68°45'42" 133°46'00"		<164m	Res
15	Gulf East Reindeer A-01	68°40'13" 134°00'31"			

* Res=Resistivity, Cry=Crystal Cable, Ac=Acoustic Velocity,
OHT=Open Hole Temperature

LOCATION NUMBER	SITE NAME	COORDINATES	0°C ISOTHERM	IBPF	METHOD
16	Gulf Ikhil I-37	68°46'34" 134°07'50"	E347	351m	Res
17	Shell Kipnik O-20	68°50'00" 134°48'19"		76m	Res
18	Shell Unak B-11	68°40'10" 135°18'40"		<158m	Res
19	Shell Ulu A-35	68°44'02" 135°52'57"	E90	<227m	Res
21	IOE Blow River Yt E-47	68°46'20" 137°27'13"		<238m	Res
22	Gulf Kamik F-38	68°57'23" 133°23'54"		<536m	Res
23	Gulf Kamik D-48	68°57'13" 133°27'30"	EX380	390m	Res
24	Gulf Kamik D-58	68°57'13" 133°29'51"		332m	Res
25	Gulf Kamik L-60	68°59'41" 133°29'24"		402m	Cry
26	Gulf Parsons F-09	68°58'28" 133°31'46"		384m	Cry
27	Gulf Parsons N-10	68°59'49" 133°31'50"	E354	332m	Res
28	Gulf Parsons N-17	68°56'53" 133°33'59"	E355	320m	Cry
29	Gulf Parsons D-20	68°59'09" 133°34'25"	E364	352m	
30	Gulf Parsons O-27	68°56'53" 133°35'56"		348m	Res
31	Gulf Parsons L-37	68°56'43" 133°39'55"	X300+	312m	Res
32	Gulf Parsons P-41	68°50'52" 133°40'29"		220m	Cry
33	Gulf Parsons L-43	68°52'38" 133°41'56"	E294	247m	Res

LOCATION		0°C			
NUMBER	SITE NAME	COORDINATES	ISOTHERM	IBPF	METHOD
34	Gulf Parsons A-44	68°53'05" 133°40'36"		352m	Res
35	Gulf Parsons P-53	68°52'49" 133°42'57"		307m	Res
36	Gulf East Reindeer G-04	68°53'16" 133°46'03"		320m	Res
37	Gulf Atigi O-48	68°57'48" 133°56'07"	EX588	564m	Cry
38	Gulf Ogruknang M-31	68°50'52" 134°24'51"		539m	Ac
39	IOE Tununuk K-10	68°59'44" 134°46'34"		137m	Res
40	Gulf Tununuk F-30	68°59'22" 134°36'43"		<166m	Res
41	Shell Tullaguk K-31	68°50'38" 135°09'22"		102m	Res
42	Shell Kugpik O-13	68°52'50" 135°18'15"	E95	<171m	Res
43	Shell Kugpik L-24	68°53'32" 135°22'13"		107m	Res
44	Dome Imnak J-29	69°08'41" 133°06'05"		533m	Res
45	Imp. Nuna A-23	69°01'14" 133°22'34"		390m	Res
46	Gulf Siku C-11	69°00'05" 133°33'49"	E378	357m	Res
47	Gulf Siku A-12	69°01'00" 133°32'31"	E360	358m	OHT
48	Gulf Siku E-21	69°00'29" 133°36'55"	394+	389m	
49	Gulf Siku C-55	69°04'04" 133°43'58"		463m	Res
50	BA IOE Reindeer D-27	69°06'05" 134°36'54"	E370	338m	

LOCATION NUMBER	SITE NAME	COORDINATES	0°C ISOTHERM	IBPF	METHOD
51	Gulf Reindeer F-36	69°05'20" 134°39'00"	EX357	340m	Res
52	Gulf Reindeer A-41	69°00'12" 134°40'19"		420m	Res
53	Gulf Kikoralok N-46	69°05'46" 134°56'33"		<161m	Res
54	Shell Titalik O-15	69°04'58" 135°03'12"		107m	Res
55	Gulf Titalik K-26	69°05'30" 135°06'15"	65+	55m	Cry
56	IMP Delta 5 Kurk M-39	69°08'55" 135°24'54"		61m	Cry
57	IOE Ellice O-14	69°03'56" 135°48'16"		38m	Cry
58	IOE Spring River Yt N-58	69°07'53" 138°44'05"		46m	OHT
59	IOE Eskimo J-07	69°16'43" 132°30'59"		201m	Res
60	IOE Tuktu O-19	69°18'55" 132°48'17"		335m	Cry
61	IOE Tuk F-18	69°17'29" 133°04'01"		<575m	Res
62	Imp. Wagnark G-12	69°11'21" 133°18'14"		387m	Res
63	Imp. Delta Wagnark C-23	69°12'01" 133°21'45"		518m	Cry
64	Gulf Red Fox P-21	69°10'48" 133°35'01"	>575	640m	Res
65	Gulf Ya Ya I-17	69°16'35" 134°32'49"		721m	Res
66	Gulf Ya Ya A-28	69°17'11" 134°35'27"	EX665	728m	Res
67	Gulf Ya Ya M-33	69°12'57" 134°39'44"		548m	Res

LOCATION NUMBER	SITE NAME	COORDINATES	0°C ISOTHERM	IBPF	METHOD
68	Gulf Ya Ya P-53	69°12'50" 134°42'45"	E433	438m	Res
69	Gulf Torpolok H-24	69°13'18" 134°50'25"		357m	Res
70	Gulf Torpolok O-54	69°13'57" 134°58'31"		216m	Res
71	Shell Kumak J-06	69°15'36" 135°00'58"		291m	Res
72	Shell Kumak K-16	69°15'33" 135°03'58"		311m	Res
73	Shell Kumak C-58	69°17'06" 135°13'54"		283m	Res
74	Shell Kumak E-58	69°17'30" 135°14'55"	E275	288m	Res
75	Shell Niglntgak B-19	69°18'11" 135°18'19"	E173	198m	Res
76	Shell Niglntgak M-19	69°18'49" 135°19'26"	>140	213m	Res
77	Shell Unipkat I-22	69°11'37" 135°20'27"	E86	88m	Res
78	Shell Niglntgak H-30	69°19'21" 135°20'35"	E144	189m	Res
79	Imp. Langley E-29	69°18'29" 135°35'56"		82m	Res
80	North Ellice J-23	69°12'34" 135°51'14"	E74	52m	Res
81	Imp. Ikattok J-17	69°16'41" 136°18'13"		<170m	Res
82	Elf et al. Kiligvak I-29	69°28'38" 131°20'16"		<159m	Res
83	Imp. Akku F-14	69°23'15" 132°19'08"		229m	Res
84	IOE Pikiolik M-26	69°25'55" 132°37'26"	362+	340m	Res

LOCATION NUMBER	SITE NAME	COORDINATES	0°C ISOTHERM	IBPF	METHOD
85	IOE Piliolik E-59	69°23'15" 132°44'35"	432+	425m	Res
87	Esso Mayogiak M-16	69°25'55" 132°49'29"		411m	Res
88	IOE Mayogiak J-17	69°26'42" 132°48'12"			
89	Imp. Mayogiak L-39	69°28'41" 132°54'30"		636m	Res
90	Gulf Kilagmiotak M-16	69°25'52" 134°04'30"		790m	Res
91	Gulf Kilagmiotak F-48	69°27'29" 134°11'51"	X600	742m	Res
92	Imp. Umiak N-10	69°29'50" 134°16'25"		768m	Res
93	Imp. Umiak J-37	69°26'36" 134°23'08"		733m	Res
94	Imp. Mallik A-06	69°25'01" 134°30'16"	>250	640m	Res
95	Imp. Mallik J-37	69°26'38" 134°38'23"		652m	Res
96	IOE Mallik L-38	69°27'44" 134°39'25"		637m	Res
97	Imp. Mallik P-59	69°28'49" 134°42'45"		707m	Res
98	IOE Taglu G-33	69°22'18" 134°53'37"			
99	IOE Taglu C-42	69°21'05" 134°56'50"	X600+	670m	Res
100	IOE Taglu D-43	69°22'14" 134°58'06"	X620	618m	Res
101	IOE Taglu H-54	69°23'20" 134°58'06"	E533	515m	Res
102	IOE Taglu D-55	69°24'14" 134°59'34"	500+	<884m	Res

LOCATION NUMBER	SITE NAME	COORDINATES	0°C ISOTHERM	IBPF	METHOD
103	IOE Taglu West P-03	69°22'55" 135°00'24"		482m	Cry
104	Chevron Upluk C-21	69°20'06" 135°21'25"		233m	Res
105	Chevron Upluk M-38	69°27'56" 135°24'54"		453m	Res
106	CCL Upluk A-42	69°21'11" 135°25'34"		378m	Res
107	Sun Garry P-04	69°23'45" 135°30'20"	502+	486m	Ac
108	Sun Garry G-07	69°26'23" 135°30'56"		422m	Res
109	Imp. Adgo C-15	69°24'13" 135°49'03"		<38m	Res
110	Imp. Adgo P-25	69°24'57" 135°50'30"	0	0m	Res
111	Esso Adgo J-27	69°26'30" 135°50'52"		<195m	Res
112	Imp. Adgo F-28	69°27'17" 135°51'16"		<151m	Res
113	Imp. Sarpik B-35	69°24'07" 135°23'10"		<38m	Res
114	Pacific Rolland Bay L-41	69°20'30" 138°56'55"		<142m	Res
115	Amaguk H-16	69°35'24" 131°02'52"		210m	Res
116	IOE Maguk A-32	69°31'09" 132°07'32"		236m	Res
117	IOE Kimik D-29	69°38'05" 132°22'10"	X663	631m	Cry
118	Imp. Atertak E-41	69°30'27" 132°42'08"	535+	553m	Res
119	Imp. Kugmallit H-59	69°38'21" 133°27'49"		696m	Res

LOCATION NUMBER	SITE NAME	COORDINATES	0°C ISOTHERM	IBPF	METHOD
120	Imp. Ivik N-17	69°36'51" 134°19'16"		697m	Res
121	Imp. Ivik J-26	69°35'42" 134°20'38"	X500	678m	Res
122	Imp. Ivik C-52	69°31'10" 134°28'52"		649m	Res
123	Imp. Ivik K-54	69°33'36" 134°29'01"		726m	Res
124	Sun Bvx Unark L-24	69°33'30" 134°37'00"		649m	Ac
125	Imp. Immerk B-48	69°37'08" 135°10'51"		542m	Res
126	Sun Pelly B-35	69°34'11" 135°23'27"		451m	Res
127	Imp. Netserk F-40	69°39'23" 135°54'21"		518m	Res
128	Imp. Netserk B-44	69°33'03" 135°55'58"		378m	Res
129	IOE Natagnak K-23	69°42'31" 131°36'44"		535m	Res
130	IOE Natagnak H-50	69°49'27" 131°40'11"		546m	Res
131	Imp. Natagnak K-53	69°42'39" 131°43'55"		576m	Res
132	IOE Atkinson H-25	69°44'20" 131°50'06"		<612m	Res
133	IOE Atkinson M-33	69°42'48" 131°54'43"		512m	Res
134	Imp. Atkinson A-55	69°44'09" 131°57'54"		619m	Cry
135	Imp. Arnak L-30	69°49'45" 133°52'21"		662m	Res
136	Imp. Pullen E-17	69°46'16" 134°19'41"		604m	Res

LOCATION NUMBER	SITE NAME	COORDINATES	0°C ISOTHERM	IBPF	METHOD
137	Imp. Nuktak C-22	69°41'07" 134°51'30"		701m	Res
138	Dome Natsek E-56	69°45'22" 139°44'35"		0m	Cry
139	Elf Horton River G-02	69°51'22" 127°15'56"		<371m	Res
140	Nicholson N-45	69°54'59" 128°56'19"		198m	Res
141	Texcan Nicholson G-56	69°55'29" 128°58'34"		378m	Res
142	Imp. Kapik J-39	69°58'32" 130°08'10"		527m	Res
143	IOE Nuvořak O-09	69°58'55" 130°30'56"		360m	Res
144	Imp. Amorok N-44	69°53'59" 130°56'16"		350m	Res
145	IOE Kanguk I-24	69°53'40" 131°05'12"		421m	Cry
146	Imp. Cigol Kanguk F-42	69°51'26" 131°11'21"		479m	Res
147	Imp. Louth K-45	69°54'32" 131°26'47"		598m	Res
148	Imp. Isserk E-27	69°56'20" 134°22'11"		674m	Res
149	Dome Tarsiut A-25	69°54'10" 136°20'20"		0m	Cry
150	Imp. Russell G-23	70°02'18" 130°06'28"		506m	Res
151	Imp. Kannerk G-42	70°01'24" 131°12'56"		665m	Res
152	Dome Ukalerk C-50	70°08'06" 132°44'08"		689m	Res
153	Dome Ukalerk 2C-50	70°09'05" 132°43'47"		690m	Res

LOCATION NUMBER	SITE NAME	COORDINATES	0°C	IBPF	METHOD
			ISOTHERM		
154	Esso Issungnak O-61	70°09'00" 134°18'48"		<90cm	Res
155	Dome Tingmiark K-91	70°10'36" 132°58'56"		552m	OHT
156	Dome Nerlerk M-98	70°27'49" 133°29'47"		668m	Res
157	Hunt Kopanoar M-13	70°22'55" 135°05'34"		402m	Cry
158	Hunt Kopanoar D-14	70°23'01" 135°05'31"			
159	Dome Kopanoar I-44	70°23'45" 135°12'00"		408m	Res
160	Dome Nektoralik K-59	70°28'35" 136°16'51"		<232m	Cl
161	Dome Kuglukik A-75	70°34'07" 130°51'22"			
162	Dome Koakoak O-22	70°21'55" 134°06'48"			

APPENDIX II

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10 ! PROGRAM TEMPMOD3 - DIANA ALLEN - AUGUST 19, 1986.
20 !
30 ! HEWLETT PACKARD SERIES 200/300 COMPUTER
40 ! HP BASIC 5.0
50 !
60 ! PROGRAM TO MODEL THE EFFECT OF
70 ! PAST CLIMATIC VARIATIONS
80 !
90 ! SOLUTION IS APPROPRIATE TO A STEP
100 ! CHANGE IN SURFACE TEMPERATURE FOR
110 ! A SEMI-INFINITE REGION Z>0
120 !
130 ! THE TEMPERATURE AT DEPTH FOR A
140 ! STEP-FUNCTION SURFACE TEMPERATURE
150 ! CHANGE IS GIVEN FOR VARIOUS TIMES
160 !
170 !  $T = T_0 + GZ + V_1 \{ \text{ERF}[Z / \{ 2 * \text{SQR}(K * T) \}] \}$ 
180 !
190 ! IDENTIFIERS
200 ! TEMPZ T FINAL TEMPERATURE AT DEPTH Z
210 ! TEMPO T0 INITIAL SURFACE TEMPERATURE
220 ! GRADIENT G GEOTHERMAL GRADIENT
230 ! DEPTH Z
240 ! CHANGET V1 CHANGE IN TEMPERATURE
250 ! NEWTEMP NEW TEMPERATURE
260 ! ERF ERROR FUNCTION
270 ! DIFF K THERMAL DIFFUSIVITY
280 ! T TIME
290 ! END OF IDENTIFIERS
300 !
310 ! OPTION BASE 1 ! array starts at (1,1)
320 ! DIM Tz(101) ! sets width of array at 101
330 ! DIM D(101) ! sets length of array at 101
340 ! DIM Ty(101) ! sets width of array at 101
350 ! DIM Dy(101) ! sets length of array at 101
360 !
370 !
380 !
390 ! PRINTER IS CRT
400 ! PRINT "THIS PROGRAM PLOTS A TEMPERATURE PROFILE AT VARIOUS TIMES"
410 ! PRINT "DUE TO A TEMPERATURE DISTURBANCE OCCURING AT TIME T=0"
420 ! PRINT "TEMPERATURES ARE CALCULATED EVERY 10 M FROM 0 TO 1000 M"
430 ! PRINT
440 ! PRINT "TO OBTAIN A PLOT OF TEMPERATURE VERSUS DEPTH RESULTING FROM"
450 ! PRINT "A STEP INCREMENT OR DECREMENT OF TEMPERATURE, ENTER THE "
460 ! PRINT "FOLLOWING PARAMETERS:"
470 ! PRINT
480 ! PRINT
490 ! INPUT "DIFFUSIVITY (IN SQ. CM/SEC)",Diff
500 ! INPUT "GEOTHERMAL GRADIENT (IN DEG/KM)",Gradient
510 ! INPUT "INITIAL SURFACE TEMPERATURE (IN DEG)",Temp0
520 ! INPUT "NEW SURFACE TEMPERATURE (IN DEG)",Newtemp
530 !
540 ! GOSUB 1340 !subroutine to construct axes
550 !
560 ! PRINTER IS CRT
570 ! PRINT "THERMAL DIFFUSIVITY (IN SQ. CM/SEC):";Diff
580 ! PRINT "GEOTHERMAL GRADIENT (IN DEG/KM):";Gradient
590 ! PRINT "INITIAL SURFACE TEMPERATURE (IN DEG C):";Temp0
600 ! PRINT "NEW SURFACE TEMPERATURE (IN DEG):";Newtemp

```

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610 PRINT "CHANGE IN TEMPERATURE (IN DEG):";(Newtemp-Temp0)
620 PRINT
630 PRINT
640 PRINT "      DEPTH (METRES) "; " TEMPERATURE (DEG C)"
650 PRINT
660 !
670 !
680 ! BEGIN TIME LOOP
690 FOR T=100 TO 1000 STEP 100
700   Pointer=1
710 ! BEGIN DEPTH LOOP
720   FOR Depth=0 TO 1000 STEP 10
730     X1=Depth/(2*SQR(Diff*1.E-4*T*3.1536E+7))
740     GOSUB 1030 ! Subroutine to calculate erf
750     V1=Newtemp-Temp0
760     Tempz=Temp0+Gradient*Depth*1.E-3+V1*(1-Erf)
770     Tz(Pointer)=Tempz ! Store Tempz in Tz array
780     D(Pointer)=Depth ! Store Depth in D array
790     Pointer=Pointer+1
800     PRINT USING "6X,5D.D,12X,3D.3D";Depth,Tempz
810   NEXT Depth
820 !
830   GOSUB 1720 ! Subroutine to plot array data
840 NEXT T
850 !
860 ! BEGIN SECOND TIME LOOP
870 FOR T=10000 TO 100000 STEP 10000
880   Pointer=1
890 ! BEGIN DEPTH LOOP
900   FOR Depth=0 TO 1000 STEP 10
910     X1=Depth/(2*SQR(Diff*1.E-4*T*3.1536E+7))
920     GOSUB 1030 ! Subroutine to calculate erf
930     Tempz=Temp0+Gradient*Depth*1.E-3+V1*(1-Erf)
940     Ty(Pointer)=Tempz ! Store Tempz in Ty array
950     Dy(Pointer)=Depth ! Store Depth in Dy array
960     Pointer=Pointer+1
970     PRINT USING "6X,5D.D,12X,3D.3D";Depth,Tempz
980   NEXT Depth
990   GOSUB 1830 ! Subroutine to plot array data
1000 NEXT T
1010 !
1020 GOTO 1890 ! End of program
1030 ! SUBROUTINE TO CALCULATE ERROR FUNCTION
1040 !
1050 ! IDENTIFIERS
1060 !   ERF          ERROR FUNCTION
1070 !   H3           SUM1
1080 !   P8           SQR(Pi)
1090 !   S3           SUM
1100 !   T5           TOL
1110 !   T4           TERM
1120 !
1130 ! GAUSSIAN ERROR FUNCTION BY INFINITE SERIES
1140 P8=SQR(3.14159) ! Square root of pi
1150 T5=1.E-3 ! Tolerance
1160 IF (X1<=0) THEN 1300
1170 IF (X1>4) THEN 1320
1180 X3=X1*X1
1190 S3=X1
1200 T4=X1

```

```

1210 I=0
1220 ! BEGIN ERF LOOP
1230 I=I+1
1240 H3=S3
1250 T4=2*T4*X3/(1+2*I)
1260 S3=T4+H3
1270 IF (T4>T5*S3) THEN 1230
1280 Erf=2*S3*EXP(-X3)/P8
1290 RETURN
1300 Erf=0
1310 RETURN
1320 Erf1=1
1330 RETURN
1340 !
1350 !
1360 ! SUBROUTINE TO CONSTRUCT AXES
1370 !
1380 GINIT                                ! graphics initialize
1390 PLOTTER IS 705,"HPGL"
1400 OUTPUT 705;"VS10"                   ! set plotter pen speed
1410 PEN 1
1420 X_gdu_max=100*MAX(1,RATIO)           ! how many gdu's wide is screen
1430 Y_gdu_max=100*MAX(1,1/RATIO)         ! how many gdu's high is screen
1440 LORG 6                               ! ref.pt : center of top of label
1450 DEG                                  ! angular mode is degrees
1460 LDIR 90                             ! specify vertical labels
1470 MOVE -.5,Y_gdu_max/2
1480 LABEL "DEPTH"
1490 LORG 4                               ! ref.pt : center of bottom
1500 LDIR 0                              ! specify horizontal labels
1510 MOVE X_gdu_max/2,.95
1520 LABEL "TEMPERATURE"
1530 VIEWPORT .1*X_gdu_max,.98*X_gdu_max,.15*Y_gdu_max,.9*Y_gdu_max
1540 WINDOW -20,15,1000,0                ! anisotropic scaling
1550 AXES 1,50,-20,0,5,2
1560 CLIP OFF                            ! labels outside viewport limits
1570 LORG 4
1580 CSIZE 4,.5                          ! smaller characters
1590 FOR I=-20 TO 15 STEP 5
1600     MOVE I,-10
1610     LABEL USING "#,K";I
1620 NEXT I
1630 LORG 8
1640 FOR J=1000 TO 0 STEP -100
1650     MOVE -20.5,J
1660     LABEL USING "#,K";J
1670 NEXT J
1680 PENUP                                ! LABEL leaves pen down
1690 RETURN                              ! end of subroutine
1700 !
1710 !
1720 ! SUBROUTINE TO PLOT ARRAY DATA
1730 !
1740 GRAPHICS ON
1750 PLOTTER IS 705,"HPGL"
1760 PEN 1
1770 Pointer=1                            ! position in array
1780 PLOT Tz(Pointer),D(Pointer)
1790 Pointer=Pointer+1
1800 IF (Pointer<102) THEN 1780

```

```
1810  PENUP
1820  RETURN
1830  Pointer=1
1840  PLOT Ty(Pointer),Dy(Pointer)
1850  Pointer=Pointer+1
1860  IF (Pointer<102) THEN 1840
1870  PENUP
1880  RETURN
1890  END
```

APPENDIX III

```

10  ! PROGRAM TEMPMOD9 - DIANA ALLEN - JUNE 25, 1987
20  !
30  ! HEWLETT PACKARD SERIES 200/300 COMPUTER
40  ! HP-BASIC 5.0
50  !
60  ! PROGRAM TO MODEL THE EFFECT OF PAST CLIMATIC VARIATIONS ON THE BASE
70  ! OF THE PERMAFROST WITH PHASE CHANGE CONDITIONS AT THE BOUNDARY.
80  !
90  ! IT IS ASSUMED THAT THE UNFROZEN REGION BELOW THE PERMAFROST REMAINS
100 ! IN A QUASI STEADY-STATE DURING THE PROCESS AND CONSEQUENTLY, THE
110 ! HEAT SUPPLIED TO THE BASE OF THE PERMAFROST AT DEPTH Z(T) REMAINS
120 ! CONSTANT AND EQUAL TO THE GEOTHERMAL FLUX Q.
130 !
140 ! HEAT BALANCE AT THE BASE REQUIRES:
150 !        $LN(dZ/dT) = Q - K1(TZ - T0)/Z(t)$ 
160 !
170 ! IDENTIFIERS
180 !       LHEAT           L           LATENT HEAT
190 !       POROSITY        N
200 !       DIFFUSIVITY     D           FROZEN THERMAL DIFFUSIVITY
210 !       CONDUCTIVITY    K1          FROZEN THERMAL CONDUCTIVITY
220 !       GRADIENT        G           GEOTHERMAL GRADIENT
230 !       HEAT_FLOW       Q           GEOTHERMAL HEAT FLOW
240 !       PERMABASE1      T0          INITIAL DEPTH OF PERMAFROST
250 !       TEMPO           T           INITIAL SURFACE TEMPERATURE
260 !       NEW_TEMP        Z(t)        NEW SURFACE TEMPERATURE
270 !       PERMABASE       t           CALCULATED PERMAFROST DEPTH
280 !       TIME_INTERVAL   TZ          TIME INTERVAL
290 !       TEMPZ           TZ          TEMPERATURE AT BASE OF PERMAFROST
300 !       CHANGE          Tz          CHANGE IN DEPTH OF PERMAFROST
310 !       TIME_CONST      Dz          TIME CONSTANT
320 !       TEMP            Tz          TEMPERATURE AT DEPTH
330 !       DEPTH           Dz
340 !       TOT_TIME
350 !       RUN_TIME        TOTAL LENGTH OF TIME INTERVALS
360 !       END OF IDENTIFIERS        DURATION OF INTERVAL
370 !
380 ! OPTION BASE 1
390 ! DIM Time(1000)           !time_constant array
400 ! DIM Changez(1000)        !change array
410 ! DIM Permabase(1000)      !new_base array
420 ! DIM Tz(101)              !temp array
430 ! DIM Dz(101)              !depth array
440 ! DIM Interval(1000)       !time_interval array
450 ! DIM Tmp(1000)            !new_temp array
460 ! DIM Base(1000)           !permabase1 array
470 !
480 !
490 ! PRINTER IS CRT
500 ! PRINT "THIS PROGRAM WILL CALCULATE THE DEPTH TO THE BASE OF THE"
510 ! PRINT "PERMAFROST RESULTING FROM SUCCESSIVE SURFACE TEMPERATURE"
520 ! PRINT "DISTURBANCES, TAKING INTO ACCOUNT THE LATENT HEAT OF FUSION"
530 ! PRINT "ABSORBED OR RELEASED AT THE FREEZING FRONT WHERE T=0 C"
540 ! PRINT "TRANSIENT THERMAL EFFECTS AT SHALLOW DEPTHS ARE ASSUMED"
550 ! PRINT "NEGLIGIBLE AND ARE NOT CONSIDERED"
560 ! PRINT "INITIAL BOUNDARY CONDITIONS INCLUDING THE INITIAL SURFACE"
570 ! PRINT "TEMPERATURE, HEAT FLOW AND THERMAL CONDUCTIVITY, MAY BE"
580 ! PRINT "ENTERED AND USED TO CALCULATE AN INITIAL STEADY-STATE"
590 ! PRINT "PERMAFROST THICKNESS - THE DEFAULT IS Z(t)-10M AT t=0"
600 ! PRINT "TEMPERATURE PLOTS AFTER EACH CHANGE IN SURFACE TEMPERATURE"

```

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610 PRINT "MAY BE REQUESTED. THESE WILL BE BASED ON SURFACE TEMPERATURE"
620 PRINT "AND THE BASE OF THE PERMAFROST AT THE END OF THE TIME INTERVAL"
630 PRINT "A LINEAR EXTRAPOLATION WILL BE MADE."
640 PRINT
650 !
660 Permabasel=0
670 Gradient=0
680 Temp0=0
690 Conductivity=0
700 !
710 INPUT "WILL YOU BE WANTING TEMPERATURE PLOTS (Y=1,N=0)",Ans1
720 IF (Ans1=1) THEN GOSUB Axes
730 !
740 INPUT "DO YOU WISH TO ENTER INITIAL BOUNDARY CONDITIONS(Y=1,N=0)",Ans2
750 IF (Ans2=1) THEN GOSUB Initial
760 !
770 IF (Permabasel>0) THEN 790
780 Permabasel=.1
790 PRINT "ENTER THE FOLLOWING PARAMETERS"
800 PRINT
810 INPUT "THERMAL DIFFUSIVITY (IN sq.cm/sec)",Diffusivity
820 INPUT "POROSITY (IN DECIMAL NOTATION)",Porosity
830 IF (Heat_flow>0) THEN 850
840 INPUT "HEAT FLOW (IN mW/sq.m)",Heat_flow
850 IF (Conductivity>0) THEN 870
860 INPUT "THERMAL CONDUCTIVITY (IN W/m/deg C)",Conductivity
870 INPUT "NEW SURFACE TEMPERATURE (IN deg C)",New_temp
880 INPUT "TIME INTERVAL (IN YEARS)",Time_interval
890 PRINTER IS PRT
900 !
910 Marker=1
920 PRINT "    BASE OF PERMAFROST (M)";"    CHANGE IN DEPTH";"    TIME"
930 !
940 ! CALCULATION OF PERMAFROST BASE
950 !
960 Pointer=1
970 Permabase(Pointer)=Permabasel
980 Time_const=(Permabase(Pointer)^2)/(4*Diffusivity*1.E-4)
990 Time(Pointer)=Time_const
1000 Change=-Time(Pointer)*(Heat_flow*1.E-3+Conductivity*New_temp/Permabase(P
ointer))/(3.33E+8*Porosity)
1010 Changez(Pointer)=Change
1020 PRINT USING "8X,5D.D,12X,5D.3D,12X,5D.D";Permabase(Pointer),Changez(Poin
ter),SUM(Time)/3.1536E+7
1030 New_base=Permabase(Pointer)+Changez(Pointer)
1040 IF ABS(Changez(Pointer))<.001 THEN 1090
1050 IF (SUM(Time)>Time_interval*3.1536E+7) THEN 1100
1060 Pointer=Pointer+1
1070 Permabase(Pointer)=New_base
1080 GOTO 980
1090 PRINT "NO FURTHER CHANGE IN PERMAFROST DEPTH"
1100 PRINT "THE FINAL PERMAFROST DEPTH (IN METRES) IS";New_base
1110 MAT Time= (0)
1120 IF (Ans1=0) THEN 1230
1130 !
1140 Number=1
1150 Gradient=-New_temp/New_base
1160 FOR Depth=0 TO 1000 STEP 10
1170 Temp=Gradient*Depth+New_temp
1180 Tz(Number)=Temp

```



```

1190      Dz(Number)=Depth
1200      Number=Number+1
1210      NEXT Depth
1220      GOSUB Plot
1230      !
1240      Interval(Marker)=Time_interval
1250      Base(Marker)=New_base
1260      Tmp(Marker)=New_temp
1270      !
1280      PRINTER IS PRT
1290      PRINT
1300      !
1310      INPUT "WOULD YOU LIKE A NEW SURFACE TEMPERATURE (Y=1,N=0)",Ans4
1320      IF (Ans4=0) THEN 1410
1330      !
1340      PRINT
1350      PRINT "YOUR INITIAL PERMAFROST DEPTH IS",New_base
1360      INPUT "WHAT IS THE NEW SURFACE TEMPERATURE (IN deg C)",New_temp
1370      INPUT "TIME INTERVAL (IN YEARS)",Time_interval
1380      Permabase1=New_base
1390      Marker=Marker+1
1400      GOTO 920
1410      !
1420      INPUT "WOULD YOU LIKE A PLOT OF THE TEMPERATURE HISTORY(Y=1,N=0)",Ans5
1430      IF (Ans5=0) THEN 1460
1440      GOSUB History
1450      !
1460      INPUT "WOULD YOU LIKE A PLOT OF THE PERMAFROST VARIATIONS(Y=1,N=0)",Ans6
1470      IF (Ans6=0) THEN 3330
1480      GOSUB Variations
1490      !
1500      GOTO 3330
1510      !
1520      Axes:      !Subroutine
1530      !
1540      ! SUBROUTINE TO CONSTRUCT AXES
1550      !
1560      GINIT                      !graphics initialize
1570      PLOTTER IS 3,"INTERNAL"
1580      PEN 1
1590      X_gdu_max=100*MAX(1,RATIO)
1600      Y_gdu_max=100*MAX(1,1/RATIO)
1610      LORG 6
1620      DEG
1630      LDIR 90
1640      MOVE -.5,Y_gdu_max/2
1650      LABEL "DEPTH"
1660      LORG 4
1670      LDIR 0
1680      MOVE X_gdu_max/2,95
1690      LABEL "TEMPERATURE"
1700      VIEWPORT .1*X_gdu_max,.98*X_gdu_max,.15*Y_gdu_max,.9*Y_gdu_max
1710      WINDOW -30,10,1000,0
1720      AXES 1,50,-30,0,5,2
1730      CLIP OFF
1740      LINE TYPE 5
1750      MOVE 0,0
1760      DRAW 0,1000
1770      LINE TYPE 1
1780      LORG 4

```

```

1790      CSIZE 4,.5
1800      FOR I=-30 TO 10 STEP 5
1810          MOVE I,-10
1820          LABEL USING "#,K";I
1830      NEXT I
1840      LORG 8
1850      FOR J=1000 TO 0 STEP -100
1860          MOVE -30.5,J
1870          LABEL USING "#,K";J
1880      NEXT J
1890      PENUP
1900      RETURN
1910 !
1920 Initial: !Subroutine
1930 !
1940 !   SUBROUTINE TO ENTER INITIAL BOUNDARY CONDITIONS
1950 !
1960      PRINT "ENTER THE FOLLOWING INITIAL CONDITIONS"
1970      PRINT
1980      INPUT "INITIAL SURFACE TEMPERATURE (IN deg C)",Temp0
1990      INPUT "HEAT FLOW (IN mW/sq.m)",Heat_flow
2000      INPUT "THERMAL CONDUCTIVITY (IN W/m/deg C)",Conductivity
2010 !
2020      Permabasel=(0-Temp0)*Conductivity/(Heat_flow*1.E-3)
2030      PRINT "THE INITIAL STEADY-STATE DEPTH OF PERMAFROST IS";Permabasel
2040      RETURN
2050 !
2060 Plot: !Subroutine
2070 !
2080 !   SUBROUTINE TO PLOT DATA
2090 !
2100      GRAPHICS ON
2110      PLOTTER IS 3,"INTERNAL"
2120      PEN 1
2130      Pointer=1
2140      PLOT Tz(Pointer),Dz(Pointer)
2150      Pointer=Pointer+1
2160      IF (Pointer<102) THEN 2140
2170      PENUP
2180      RETURN
2190 !
2200 History: !Subroutine
2210 !
2220 !   SUBROUTINE TO PLOT TEMPERATURE HISTORY
2230 !
2240      PRINTER IS CRT
2250      GCLEAR
2260      PRINT "CHANGE PAPER IN PLOTTER"
2270      PRINT
2280      PRINT "WHEN PLOTTER IS READY, PRESS 'CONTINUE'"
2290      PAUSE
2300      GINIT
2310      GRAPHICS ON
2320      PLOTTER IS 3,"INTERNAL"
2330      PEN 1
2340      X_gdu_max=100*MAX(1,RATIO)
2350      Y_gdu_max=100*MAX(1,1/RATIO)
2360      LORG 6
2370      DEG
2380      LDIR 90

```

```

2390  MOVE -.5,Y_gdu_max/2
2400  LABEL "TEMPERATURE"
2410  LORG 4
2420  LDIR 0
2430  MOVE X_gdu_max/2,95
2440  LABEL "TIME (YEARSx10000)"
2450  VIEWPORT .1*X_gdu_max,.98*X_gdu_max,.15*Y_gdu_max,.9*Y_gdu_max
2460  WINDOW 0,15,-20,5
2470  AXES .5,1,0,5,2,5
2480  CLIP OFF
2490  LORG 4
2500  CSIZE 4,.5
2510  FOR I=0 TO 15 STEP 1
2520      MOVE I,5.4
2530      LABEL USING "#,K";I
2540  NEXT I
2550  LORG 8
2560  FOR J=-20 TO 5 STEP 5
2570      MOVE -.1,J
2580      LABEL USING "#,K";J
2590  NEXT J
2600  MOVE 0,0
2610  DRAW 15,0
2620  PENUP
2630  !
2640  PEN 4
2650  Tot_time=SUM(Interval)
2660  Number=1
2670  PLOT Tot_time/10000,Tmp(Number)
2680  Run_time=Tot_time-Interval(Number)
2690  PLOT Run_time/10000,Tmp(Number)
2700  Number=Number+1
2710  PLOT Run_time/10000,Tmp(Number)
2720  Run_time=Run_time-Interval(Number)
2730  IF (Run_time=0) THEN 2750
2740  GOTO 2690
2750  PLOT Run_time,Tmp(Number)
2760  RETURN
2770  !
2780  Variations:      !Subroutine
2790  !
2800  ! SUBROUTINE TO PLOT PERMAFROST VARIATIONS
2810  !
2820  GCLEAR
2830  PRINT "CHANGE PAPER IN PLOTTER"
2840  PRINT
2850  PRINT "WHEN PLOTTER IS READY, PRESS 'CONTINUE'"
2860  PAUSE
2870  GINIT
2880  GRAPHICS ON
2890  PLOTTER IS 3,"INTERNAL"
2900  PEN 1
2910  X_gdu_max=100*MAX(1,RATIO)
2920  Y_gdu_max=100*MAX(1,1/RATIO)
2930  LORG 6
2940  DEG
2950  LDIR 90
2960  MOVE .5,Y_gdu_max/2
2970  LABEL "BASE OF PERMAFROST"
2980  LORG 4

```

```
2990  LDIR 0
3000  MOVE X_gdu_max/2,95
3010  LABEL "TIME (YEARSx10000)"
3020  VIEWPORT .1*X_gdu_max,.98*X_gdu_max,.15*Y_gdu_max,.9*Y_gdu_max
3030  WINDOW 0,15,1000,0
3040  AXES .5,50,0,0,2,2
3050  CLIP OFF
3060  LORG 4
3070  CSIZE 4,.5
3080  FOR I=0 TO 15 STEP 1
3090      MOVE I,-10
3100      LABEL USING "#,K";I
3110  NEXT I
3120  LORG 8
3130  FOR J=1000 TO 0 STEP -100
3140      MOVE -.2,J
3150      LABEL USING "#,K";J
3160  NEXT J
3170  !
3180  PEN 5
3190  Tot_time=SUM(Interval)
3200  Number=1
3210  LINE TYPE 5
3220  PLOT Tot_time/10000,Base(Number)
3230  Run_time=Tot_time-Interval(Number)
3240  PLOT Run_time/10000,Base(Number)
3250  LINE TYPE 1
3260  Number=Number+1
3270  Sum_time=Run_time-Interval(Number)
3280  PLOT Sum_time/10000,Base(Number)
3290  IF (Sum_time=0) THEN 3320
3300  Run_time=Sum_time
3310  GOTO 3260
3320  RETURN
3330  END
```

END

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FIN