GEOLOGICAL SURVEY OF CANADA
COMMISSION GÉOLOGIQUE DU CANADA

PAPER 75-28

PERMAFROST IN THE MACKENZIE DELTA, NORTHWEST TERRITORIES

MICHAEL W. SMITH

Energy, Mines and Resources Canada
Énergie, Mines et Ressources Canada

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Catalogue No. M44-75-28

Price subject to change without notice

Information Canada
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1976

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Variations in ground temperature regime and permafrost distribution were studied between 1969 and 1971 in a small area in the east-central part of the Mackenzie Delta about 50 km northwest of Inuvik, Northwest Territories. The Delta is an area of active sedimentation and erosion, and about 50 per cent of the study area is covered by water. A major distributary in the study area is undergoing lateral migration, and the local configuration of permafrost is closely related to the history of river migration. The vegetation shows a successional sequence related to river migration indicating a complex interaction between vegetation, topography, and microclimate.

The major objectives of the study were to:

(1) describe the permafrost and ground temperature variations in the study area;
(2) understand how local environmental factors influence the ground temperature field; and
(3) analyze the development of the present ground temperature field in terms of its position in the long-term migration history of a shifting channel.

In order to monitor the ground temperature variations, boreholes were drilled to various depths up to 30 m. Temperatures were measured with thermistors, and these measurements were augmented with seismic and resistivity surveys. Lake and river temperatures also were recorded and ground materials were sampled from boreholes. Measurements of ground temperatures, snow depths, and ice cover were made in winter.

Permafrost is discontinuous; calculations indicate that permafrost is absent beneath channels and lakes that are wider than 80 to 100 m. Ground temperatures are warmer close to water bodies, and the permafrost table falls steeply beneath cutbanks. Observations indicate permafrost thicknesses of between 50 to 65 m in stable, spruce-covered areas. Calculations show that the maximum permafrost thickness in the area is about 100 m at sites most distant from water bodies. Beneath slip-off slopes, where temperatures are warmer than beneath cutbanks, permafrost is only 2. 5 to 9 m thick, thickens away from the river, and disappears towards it. Permafrost is absent in some places where winter snow drifts are deep.

Using simple heat conduction theory a consistent explanation of permafrost distribution in terms of local environmental factors is developed. The heat conduction models are suitable for ground temperature prediction, with agreement typically within ±0. 5°C of observed values for most sites. Calculations of thermal disturbance due to channel shifting are in general agreement with observations, although omission of the latent heat term leads to some errors.

Résumé

On a étudié les variations du régime de température du sol et la répartition du pergélisol, entre 1969 et 1971, dans une petite zone du centre est du delta du Mackenzie, à environ 50 km au nord-ouest d'Inuvik, dans les Territoires du Nord-Ouest. Le delta est une région active de sédimentation et d'érosion et près de 50 pour cent de la région étudiée est couverte d'eau. Un bras majeur de ce delta qui se situe dans la région étudiée se déplace latéralement et la configuration locale du pergélisol est étroitement liée à la migration de ce difluent. La végétation se répartit en associations successives reliées à la migration de ce bras, ce qui démontre une interaction complexe entre la végétation, la topographie et le microclimat.

Voici quels ont été les grands objectifs de l'étude:

(1) décrire le pergélisol et les variations de température du sol dans la zone étudiée;
(2) comprendre comment les facteurs locaux de l'environnement influent sur le champ thermique du sol; et
(3) analyser l'évolution du champ thermique actuel du sol quant à sa position dans l'histoire des migrations à long terme d'un chenal en voie de déplacement.

Afin de surveiller les variations de températures de sol, on a foré des trous de sonde à diverses profondeurs allant jusqu'à 30 mètres. On a mesuré les températures avec des thermistances, et on a procédé, de plus, à des levés sismiques et de résistivité. On a enregistré également les températures des lacs et des rivières et l'on a prélevé des échantillons du sol dans les trous de sonde. En hiver, on a mesuré les températures du sol, la profondeur de la neige et la glace de couverture.

Le pergélisol n'est pas continu: des calculs montrent qu'il n'y a pas de pergélisol sous les chenaux et les lacs de plus de 80 à 100 m de large. Les températures du sol sont plus chaudes près des nappes d'eau et le niveau du pergélisol tombe brusquement sous les berges hautes. On a observé que l'épaisseur du pergélisol varie de 50 à 65 m dans les zones stables et recouvertes d'épinettes. On a déterminé par calcul que l'épaisseur maximale du pergélisol dans la zone étudiée est de 100 m aux endroits les plus éloignés des nappes d'eau. Sous les versants de glissement, où la température est plus chaude que sous les berges hautes, le pergélisol n'a une épaisseur que de 2. 5 à 9 m et il s'épaissit au fur et à mesure qu'il s'éloigne de la rivière pour disparaître quand il s'en approche. Il n'y a pas de pergélisol en certains endroits où les congères accumulées en hiver sont très épaisses.

L'auteur, à l'aide d'une simple théorie de la transmission de la chaleur, élabore une explication cohérente de la répartition du pergélisol selon l'évolution de facteurs de l'environnement local. Les modèles de transmission de la chaleur conviennent à la prévision de la température des sols et la marge d'erreur est de -0. 5°C pour la plupart des endroits observés. Les calculs de perturbation thermique due au déplacement des chenaux concordent généralement avec les observations, bien que l'omission du terme de chaleur latente entraîne certaines erreurs.

LIST OF SYMBOLS

T	temperature (^oC) ($\bar{T}$ indicates term mean)
T_s	surface temperature
T_d	temperature of the disturbance (e.g. river temperature)
T_w	water temperature
x, y	horizontal co-ordinates (m)
z	depth (m) ($z = 0$ is the surface and z is positive downwards)
t	time since initiation of temperature disturbance (days, years)
α	thermal diffusivity (m^2 $year^{-1}$)
k	thermal conductivity (W cm^{-1} $^oC^{-1}$)
C	volumetric heat capacity (J cm^{-3})
$\theta(x, y, z, t)$	the thermal disturbance produced by T_d on the temperature at any (x, y, z) after any time t (^oC)
P	period of the disturbance wave
v	migration speed
x_c	tautochrone crossing point
s	width of strip (river)
Gg	earth's geothermal gradient (^oC m^{-1})
Ww	total soil moisture content on a dry weight basis
erfc x	$= \frac{2}{\sqrt{\pi}} \int_x^\infty e^{-u^2}\, du$

PERMAFROST IN THE MACKENZIE DELTA, N.W.T.

INTRODUCTION

Many studies have been concerned with the qualitative effect of differences in climate, vegetation, topography, lithology, and hydrology on the distribution of permafrost. Although obviously basic to a general understanding of the problem, such an approach cannot provide specific information on the local configuration of permafrost. The quantitative aspects of the thermal relationships between the ground temperature field and physical factors, and the details of permafrost configuration, can be resolved only by geothermal investigations. Measurements of ground temperature, at the surface and at depth, and measurements of thermal properties are needed to determine the history of permafrost and the degree to which its configuration reflects present boundary conditions.

To understand thoroughly any geothermal studies there must be a correlation with the properties of the ground materials, the surface cover, topographic position, past history, and present climatic conditions. Analysis of temperatures and their gradients may then make it possible to determine whether permafrost at any locality is aggrading or degrading, and in turn may provide basic information on the physical environment of the present and perhaps of the past.

An important and intriguing problem, both from a scientific and engineering viewpoint, is to determine the disturbance of subsurface temperatures that result when the temperature at the ground surface within some finite area differs from the surface temperature characteristic of the region at large. Such conditions might relate to the presence of natural features such as lakes and rivers; an area with different surface thermal properties; or modifications of the surface as a result of erecting buildings, vegetation removal, highway construction and pipelines. The application of heat conduction theory to this general class of problems has been treated by Lachenbruch (1957a, 1957b, 1959) and W.G. Brown (1963a, 1963b). A current major objective of applied permafrost research is the development of heat flow models designed to predict the effects of introduced sources and sinks, such as pipelines, on the thermal regime (Lachenbruch, 1970; Hwang *et al.*, 1972). Numerical models must be used when irregularities of problem geometry or initial boundary conditions make any analytic solution impossible.

Permafrost and Heat Conduction Theory

Valuable contributions from Lachenbruch (1957a, 1957b, 1959, 1970), Lachenbruch *et al.* (1962), Mackay (1962, 1963, 1971), W.G. Brown (1963a, 1963b), and W.G. Brown *et al.* (1964) have shown that physical theory developed in the field of heat conduction might profitably be applied to the analysis of the ground thermal regime in the natural physical environment.

Since groundwater in permafrost regions generally is immobilized as ice so that only localized groundwater circulation can occur, Lachenbruch *et al.* (1962, p. 792) assumed that for the most part permafrost temperatures are determined almost entirely by conductive transfer, and that heat conduction models can be used with some confidence. Russian workers have called attention to the role of infiltration and groundwater circulation in the development of taliks beneath water bodies (Efimov, 1964; Mel'nikov, 1964; Romanovskii and Chizhov, 1967). Dostovalov and Kudryavtsev (1968, p. 319-321) discuss, in general terms, the degradation of permafrost caused by groundwater flow through fractures and fissures in rock. Tiutiunov (1961) and Leschikov and Zarubin (1967) describe field situations where taliks have been formed in bedrock by this process.

The Present Study

The major objectives of this study were to:

(1) determine the variations in ground temperatures and permafrost distribution in the study area;
(2) understand how local physical factors influence permafrost distribution; and
(3) analyze the development of the present ground temperature field in terms of its position in a long-term geomorphological sequence (river migration).

These objectives are embodied in the attempt to:

(1) demonstrate the consistency of the ground temperature field through the framework of simple heat conduction theory; and thereby
(2) derive a predictive model for ground temperatures and permafrost distribution. Successful heat flow models then can be applied to areas for which no permafrost observations are available.

Geomorphological and biological evidence shows that the land surface conditions in the study area are changing with time. Relationships obtained for the present set of environmental conditions must be interpreted in terms of the degradation and re-establishment of permafrost under these changing boundary conditions.

In the application of simple heat conduction theory to this problem it was necessary to neglect the energetics of latent heat. There is no analytical solution to the two-dimensional phase change problem. The one-dimensional thawing or freezing problem has been treated extensively in the literature (*see* Muehlbauer and Sunderland, 1965); however, there are many problems that are distinctly two or three dimensional and cannot be approximated by a one-dimensional analysis.

Original manuscript submitted: March 29, 1974
Final version approved for publication: July 15, 1975
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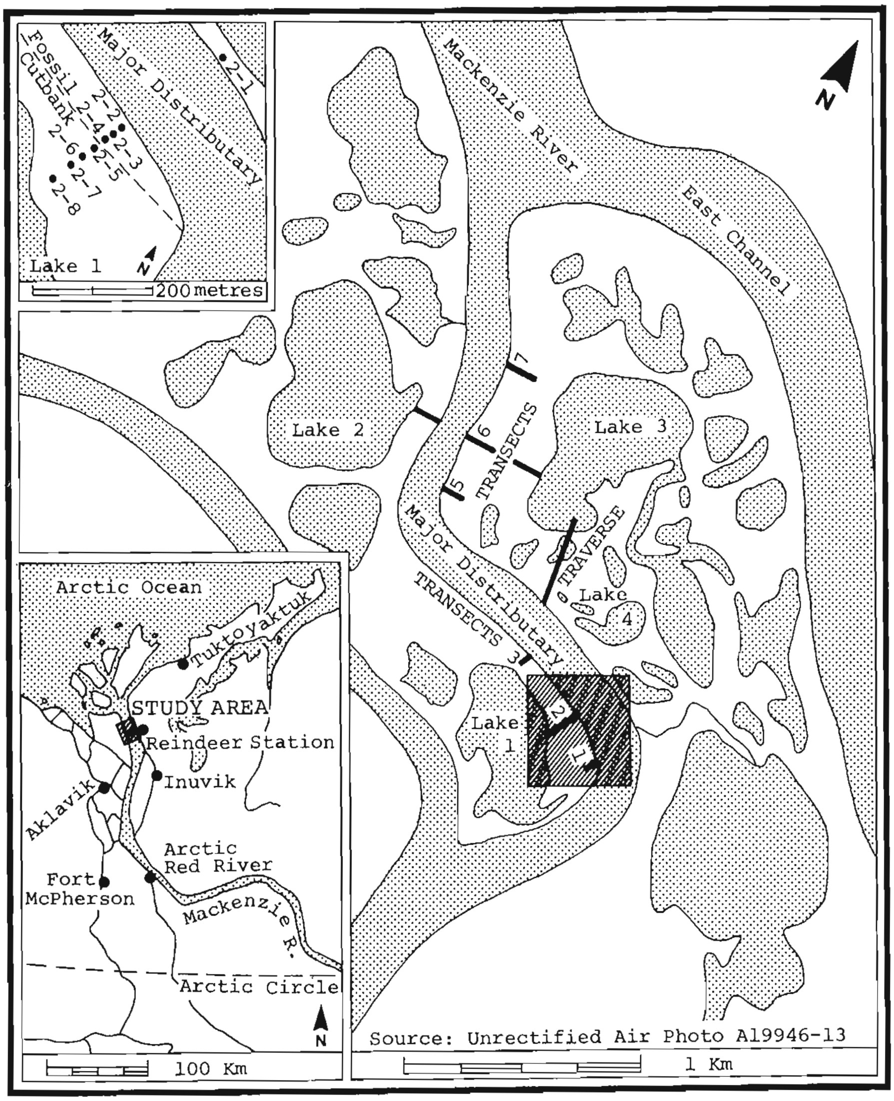

Figure 1.

The study area.

The models used in this study have the advantages of being universally available and of successfully accounting for the problem geometry. Neglect of latent heat effects, though, does lead to some distortion in the shape and rate of penetration of isotherms. It is felt, however, that such models can provide much useful information about ground temperatures and permafrost distribution, since predictions from these simple models generally do agree with observed values.

The models presented here are not appropriate for detailed thermal analyses of engineering projects where it is necessary to consider latent heat effects and perhaps convective heat transfer.

Some Definitions

Permafrost is defined exclusively on the basis of temperature and is interpreted in this study to include rock or soil material, with or without included moisture, that has remained below 0°C for some period of time--the minimum period being defined by R.J.E. Brown (1969) as including ground that freezes in one winter, remains frozen through the following summer and into the next winter. This definition, therefore, includes the pereletok[1] of other authors.

The words "freezing" and "thawing", commonly referring to the change of state between water and ice, will be used here as if these processes actually take place at 0°C. Permafrost may be ice-free below 0°C if the moisture it contains is saline or if it contains no moisture at all. In most soils, and especially those with a clay fraction, substantial amounts of water, in the liquid phase, can persist at temperatures below 0°C (P.J. Williams, 1967). Ice in permafrost can occur as coatings, grains, veinlets, or massive beds (Mackay, 1972).

The Thermal Effects of Water Bodies

The thermal effects of water bodies in high latitudes constitute the greatest local departure of the ground temperature field from the systematic geographical patterns determined by climatic factors (Lachenbruch *et al.*, 1962). Mean annual temperatures beneath water

[1]A glossary of terms is presented in Appendix.

bodies are commonly anomalously high; this is associated with the high heat capacity of water, combined with the reduction in the winter loss of heat due to an ice and snow cover. The Mackenzie Delta is icebound for almost eight months of the year.

The influence of water bodies on permafrost configuration is commonly quite dramatic; Hopkins *et al.* (1955) reported that permafrost is absent or lies at great depths beneath lakes and ponds throughout Alaska. If a freshwater body is deeper than the maximum thickness of winter ice, its bottom sediments will have a mean annual temperature of greater than $0^{o}C$. In high latitudes, where the mean annual ground surface temperature may be -10^{o} to $-15^{o}C$, lakes more than 2 m deep generally do not freeze to the bottom (Brewer, 1958b; Johnston and Brown, 1964; Inuvik Research Laboratory, 1968). Beneath even relatively shallow bodies of water, depression of the permafrost table occurs (e. g. Brewer, 1958a; Grigor'ev, 1959; Vturina, 1960; Johnston and Brown, 1964; Anisimova, 1966).

Thermal effects associated with rivers should be similar to those of lakes, except that water flowing in from elsewhere will be important in determining the mean annual river temperature. Virtually no temperature data are available from beneath rivers in permafrost regions, and most of the information on permafrost occurrence in these situations is of a qualitative nature, inferred mainly from drilling operations for groundwater and mining exploration. Further, most of the information pertains to conditions along river banks or on flood plains, although it does appear to confirm the presence of taliks beneath the rivers themselves (Wallace, 1948; Cederstrom, 1950; Fernald, 1960; J.R. Williams, 1970). Where the river freezes through in winter there may still be a warming effect but a talik as such will not exist (Samson and Tordon, 1969).

In the U. S. S. R. unfrozen zones are known to exist beneath large lakes and rivers (Svetozarov, 1934; Grigor'ev, 1959; Vturina, 1960; Efimov, 1964; Mel'nikov 1964; Anisimova, 1966; Nekrasov, 1967; Romanovskii, 1967).

Permafrost Configuration and River Migration

Migration of a river across the land surface can be expected to cause migration of the talik beneath the river. The thermal effects of the river on the surrounding ground temperature field depend not only on the strength of this source, but also upon the length of time available to the thermal processes, which is a function of the speed of migration. The local configuration of permafrost, therefore, is closely related to the history of river migration.

Péwé (1965) described the distribution of permafrost beneath a slip-off slope on the Yukon River. He found that the frozen alluvium forms a wedge-shaped mass that is thin near the river but thickens away from it. J.R. Williams (1970) described the distribution of permafrost in the flood plains of the Tanana and Chena rivers near Fairbanks. Based on data from over 5100 wells and borings, he summarized the pattern as a progressive increase in permafrost thickness from the flood plain to successively older terraces, suggesting that permafrost thickness and continuity are partly a function of the time elapsed since removal of the warming influence of the river (J.R. Williams, 1970, p. 37).

PERMAFROST IN THE MACKENZIE DELTA

Introduction

The Mackenzie Delta is a low, flat area, covering more than 15 000 km^2, spotted with thousands of lakes and dissected by an intricate anastomosing network of several large channels and numerous smaller winding channels. The channels rarely meander, in the geomorphic sense, although they may be sinuous and they do wander (Mackay, 1963, p. 105).

The present study was carried out between 1969 and 1971 in the east-central part of the modern Mackenzie Delta, some 50 km northwest of Inuvik, encompassing an area of about 10 km^2 (Fig. 1). Aspects of the physical geography of the region have been described by Mackay (1963), whilst the vegetation and geomorphology in the study area have been investigated by Gill (1971). The climate of this part of the Delta is tansitional between arctic and subarctic. Some relevant climatic data are presented in Table 1.

Previous Investigations

R.J.E. Brown (1967) uses the $17^{o}F$ ($-8.3^{o}C$) mean annual air isotherm to delimit the region of continuous permafrost. Aklavik and Inuvik have mean annual air temperatures below $17^{o}F$ (Table 1). If it were to conform to the broad geographical patterns of permafrost distribution, the region would be included in the continuous zone. Based on ground temperature, however, the region is seen to form an outlier of the discontinuous permafrost zone; Mackay (1967) quotes temperatures of between -2.4^{o} to $-3.8^{o}C$ (at a depth of 15.2 m) for various Delta locations. His data also indicate that Arctic Red River is close to the boundary (a temperature of between -4^{o} and $-5^{o}C$ at 15.2 m), but that Fort McPherson should properly be classified as discontinuous (a temperature of about $-3^{o}C$ at 15.2 m).

At Aklavik several piles were driven into the bottom of Peel Channel and no permafrost was found (R.J.E. Brown, 1956). Investigations by Johnston and Brown (1964) revealed no permafrost beneath a small, shallow Delta lake; the thawing effect was confined to the ground beneath the lake, as indicated by the presence of permafrost at the shoreline.

The Mackenzie Delta is thus marginal between the two permafrost zones. Permafrost is physically discontinuous, and thicknesses are probably generally less than 100 m (Johnston and Brown, 1964). This pattern is undoubtedly due to the thermal influence of the large amounts of surface water which occupy up to 50 per cent of the surface area in some parts of the Delta (Mackay, 1963, p. 98). On nearby tundra, just a kilometre or so from the Delta, a permafrost thickness of about 365 m has been measured (Jessop, 1970).

TABLE 1

Averages and Extremes of Climatic Data

Inuvik: 68°18'N, 133°29'W; Elevation 67 m; Period 1957-1971

	TEMPERATURE (°C)					PRECIPITATION (cm)	
	Mean daily	Mean daily maximum	Mean daily minimum	Extreme maximum	Extreme minimum	Mean total water	Mean snow on ground
Jan	-31.1	-26.0	-36.1	1.7	-51.7	2.0	58.8
Feb	-29.8	-24.4	-35.2	2.8	-56.7	1.2	60.0
Mar	-24.7	-18.3	-31.1	6.1	-50.6	1.5	65.5
Apr	-14.5	- 7.9	-21.1	13.3	-46.1	1.6	56.1
May	- 0.8	4.2	- 5.3	23.3	-27.8	1.7	1.3
June	10.2	16.3	4.1	31.7	- 6.1	1.9	0.0
July	13.4	19.3	7.6	31.1	- 3.3	3.7	0.0
Aug	10.4	15.6	5.1	29.4	- 6.1	4.6	0.0
Sept	2.8	6.9	- 1.4	25.6	-18.9	2.1	0.7
Oct	- 8.0	- 4.4	-11.6	15.0	-35.0	3.7	25.0
Nov	-21.6	-17.1	-26.2	10.6	-46.1	1.7	33.6
Dec	-26.9	-21.9	-31.9	- 0.6	-50.0	1.8	46.6
Year	-10.1					27.5	

Aklavik: 68°14'N, 135°00'W; Elevation 9 m; Period 1931-1960

	TEMPERATURE (°C)					PRECIPITATION (cm)
	Mean daily	Mean daily maximum	Mean daily minimum	Extreme maximum	Extreme minimum	Mean total water
Jan	-28.8	-24.9	-32.7	6.7	-50.6	1.4
Feb	-27.6	-23.6	-31.6	9.4	-52.2	1.0
Mar	-22.6	-17.6	-27.5	9.4	-48.9	1.0
Apr	-12.7	- 7.1	-18.3	13.9	-42.2	1.1
May	- 0.7	3.9	- 5.3	25.0	-25.6	0.9
June	9.4	14.3	4.5	30.0	- 6.7	1.9
July	13.6	18.3	8.9	33.9	- 1.1	3.2
Aug	10.4	14.6	6.1	31.1	- 3.9	3.5
Sept	3.4	6.4	0.3	24.4	-11.1	2.8
Oct	- 7.0	- 4.4	- 9.6	12.8	-30.0	2.6
Nov	-19.9	-16.5	-23.3	6.7	-45.6	1.8
Dec	-27.0	-23.2	-30.8	10.0	-47.8	1.2
Year	- 9.1					22.4

The Study Area

The Mackenzie Delta is an area of active sedimentation and erosion. Shifting channel courses, as evidenced by abandoned channels, point bar deposits, and undercut banks, are a conspicuous element of the landscape. Local relief generally does not exceed 3 to 4 m, excluding channel cross-sections; between levee systems relief is generally minor. Vegetation shows a sequential distribution. The actively forming sections adjacent to the channels are bare of vegetation; with increasing distance from the river willow (*Salix* spp.) and alder (*Alnus crispa*) grow. Behind this vegetation is the inactive part of the flood plain which is populated by spruce (*Picea glauca*), the typical climax community in the lower valley and delta of the Mackenzie River.

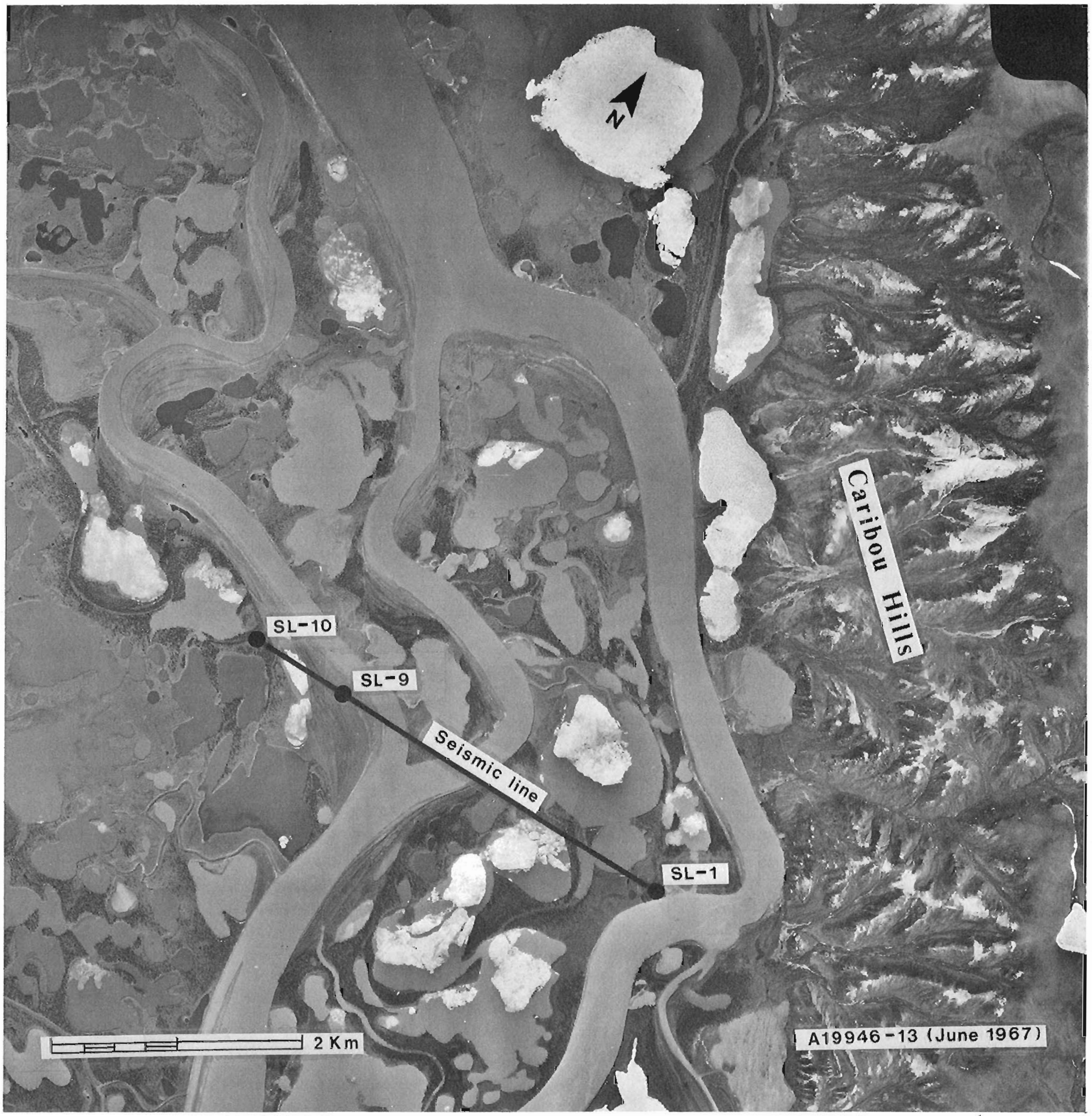

Figure 2. Study area and vicinity (portion of air photograph A19946-13).

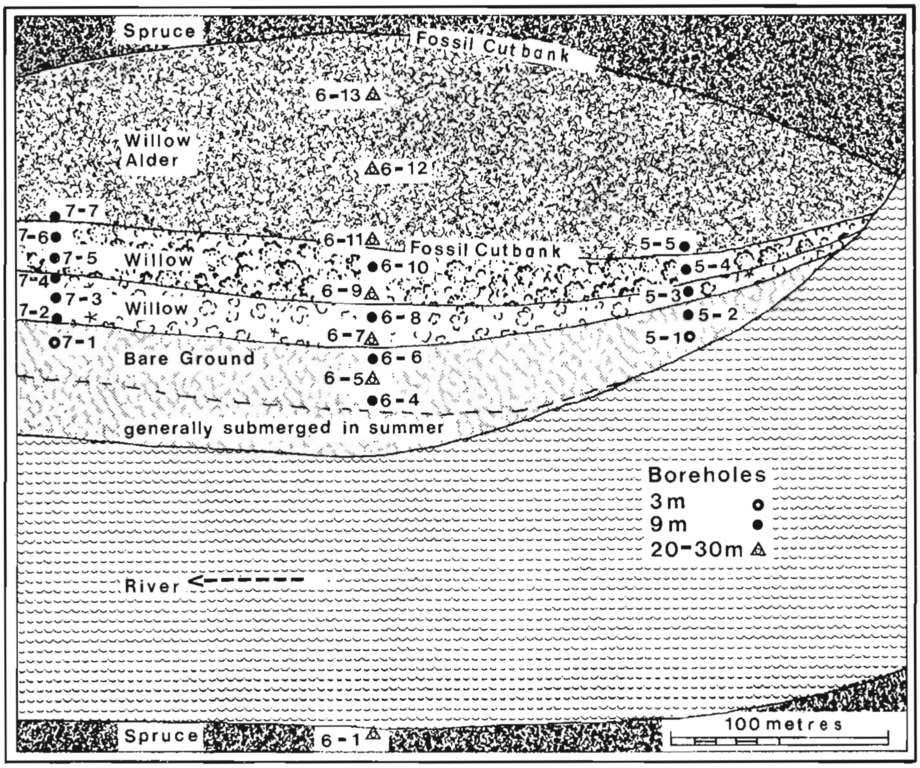

Figure 3.

Temperature borehole network on a slip-off slope.

The study area contains many lakes, which range in size from 0. 01 km^2 up to 0. 5 km^2 (Fig. 2). About 50 per cent of the area is covered by water bodies. Lakes sounded were found to have maximum depths of 1. 5 to 2. 5 m (at low water) which is greater than the winter ice thickness.[1] In March 1971 ice thicknesses varied from 0. 85 to 1. 0 m; in March 1970 they were about 15 cm thinner; and in March 1969 they were about 20 to 25 cm thicker. Maximum ice thicknesses on four Delta lakes measured in 1965-66 ranged between 1. 0 and 1. 2 m (Inuvik Research Laboratory, 1968). These ice thicknesses are less than those reported from the Arctic Coastal Plain; lower temperatures and less snowfall in that region may be the reason.

The major distributary, 120 to 170 m wide, flows through the area, is erosionally active, and exhibits lateral migration (Fig. 2). The channel bottom profile is fairly smooth and broadly U-shaped with an average maximum depth of about 4 to 6 m. In March 1971 the ice thickness varied from 1. 0 to 1. 3 m. On the outside bends of meanders the river is cutting into a mature, spruce-covered surface more than 300 years old (as indicated by tree cores), with consequent degradation of permafrost. The undercut slope is marked by a levee which maintains its altitude and backslope as it erodes back. As the cutbank recedes new deposits are formed on the slip-off slope, and under low mean annual surface temperatures permafrost will form there again.

These geomorphic changes are accompanied by a vegetation succession which produces a complex interaction between topography, vegetation, and microclimate and the ground thermal regime. This vegetation gradient introduces other sources of variation in ground temperatures. It is possible to identify various terrain segments; because of the close association between vegetation and topographic location they are termed: a) bare ground; b) willow (1) (snow bank zone); c) willow (2); d) willow-alder; and e) spruce (Fig. 3). Gill (1971) gives a full description of the relationship between vegetation and topography in this area.

The modern Delta is thought to have aggraded at a mean rate of about 5 mm per year for the past 7000 years or so (Johnston and Brown, 1965). According to Ritchie and Hare (1971) it is likely that climatic conditions conducive to the existence of permafrost have persisted for at least 5500 years. If, however, it is assumed that permafrost is totally destroyed by the presence of a river above it, then the permafrost at any location will only be as old as the time elapsed since a river, or lake, last occupied the surface above it (P.J. Williams, 1968, p. 1386). Thus the changing configuration of permafrost is compatible with the time scale of geomorphic change, and there need be no concern with any processes on a longer time scale.

Field Data

Ground Temperatures

Borehole temperatures were measured with thermistors which were sealed at intervals along multiconductor cable. The pods containing the thermistors

[1] Depths in excess of the winter ice thickness must be consistently present in a large number of Delta lakes, since the region is an important habitat for the muskrat, which lives in lakes with unfrozen pools (Mackay, 1963, p. 135).

TABLE 2

River and Lake Temperature Data

Month	MONTHLY MEAN River temperatures (°C)				MONTHLY MEAN Lake temperatures (°C)	
	1967*	1968**	1969**	1970	1968**	1969**
Jan	(0.0)[1]	(0.0)	(0.0)	(0.0)	(0.0)[1]	(0.0)
Feb	(0.0)	(0.0)	(0.0)	(0.0)	(0.0)	(0.0)
Mar	(0.0)	(0.0)	(0.0)	(0.0)	(0.0)	(0.0)
Apr	(0.0)	(0.0)	(0.0)	(0.0)	(0.0)	(0.0)
May	0.2	(0.2)	0.2	0.2	0.2	(0.2)
June	8.4	9.2	10.2	9.2	(7.2)	7.0
July	14.2	15.6	14.8	16.3[2]	15.6	14.2
Aug	12.6	14.2	10.5	15.0	12.7	8.1
Sept	7.3	(7.6)	7.9	(7.6)	(5.4)	5.4
Oct	(1.5)	(1.5)	1.5[3]	(1.5)	(1.0)	(1.0)
Nov	(0.1)	(0.1)	(0.1)	(0.1)	(0.1)	(0.1)
Dec	(0.0)	(0.0)	(0.0)	(0.0)	(0.0)	(0.0)
Year	3.7	4.0	3.8	4.2	3.5	3.0

[1]Values in parentheses are estimates based on incomplete or limited field data from other years. The following temperature data refer to various winter dates:

	March 1969 (°C)	March 1970 (°C)	December 1970 (°C)
River water	0.0	0.05	0.0
River bottom	0.0	0.07	0.0
Lake water	0.0	0.08	-
Lake bottom	0.0	0.08	-

[2]The following temperature soundings were made on July 24, 1970 (p.m.) and indicate that the river is well mixed to a uniform temperature:

	Depth (m)	Temp (°C)		Depth (m)	Temp (°C)
Sounding no. 1	1.0	16.7	Sounding no. 2	1.0	16.7
	2.0	16.6		2.0	16.7
	3.0	16.6		3.0	16.6
	4.0	16.7		4.0	16.6
	4.3 (bottom)	16.8		5.0	16.6
				6.0	16.8
				6.5 (bottom)	16.8

[3]Estimate based on measurements for October 1 to 6 only.

* Data supplied by D. Gill (pers. comm.).

** Data supplied by C.P. Lewis (pers. comm.).

were covered with rubber sealant and then with self-vulcanizing rubber splicing tape. Temperatures were mostly read manually using a simple bridge circuit; the absolute accuracy with this type of thermistor is ±0.1°C.[1] Any individual thermistor could be read easily with a resolution of ±0.02°C, but slight variations in calibration between thermistors reduce overall accuracy. On some occasions, temperatures were recorded on a chart recorder but resolution was not as good as with the manual measurements (±0.1°C). The recorders were calibrated against a precision decade resistance, and overall accuracy probably was maintained at ±0.25°C.

Only a limited number of boreholes could be drilled and thus there was a problem in obtaining an adequate and representative spatial sample. The geomorphic sequence previously described was selected as a focus in the main study area, and major borehole transects were located optimally along it so that all terrain segments were well represented (Figs. 1 and 3).

Boreholes were drilled with a 10-h.p. Winkie Drill using a three-wing drag bit, giving a hole about 5 to 6 cm in diameter. An auxiliary pump was used to circulate fluid during drilling. In 1969 a calcium chloride solution was used in order to prevent the possibility of freezing in the hole during drilling. Unfortunately this caused the magnesium-zirconium rods to corrode and subsequently fracture. In 1970 ordinary lake and river water was used without any operating problems. The drilling period for a 30-m hole was only four or five hours. With drilling, the thermal regime is disturbed, and a period of two to three weeks is required for the temperature to return to equilibrium.

It was not possible to drill under water bodies, but holes were located as close as possible to the channel. Fifty-four boreholes were drilled (1969-1970) to various depths up to 30 m, which proved to be the deepest practical. More than 300 thermistors were placed in boreholes. The network design is illustrated in Figure 3.

Ground Materials

Since the cuttings were flushed to the surface by pumped water, the drilling procedure was unsuitable for sampling ground materials. In March 1970, however, a seismic line was located in the vicinity of the study area (Fig. 2) which provided an opportunity to obtain samples. Through the co-operation of Imperial Oil Limited and Gulf Oil Limited ten boreholes were drilled, the deepest to 32 m. Drilling was performed using compressed air, and it was possible to collect 3-m integrated samples over the depth of each borehole. These samples subsequently were analyzed for total moisture content, grain-size characteristics, and organic content.

Other Data

The knowledge of permafrost distribution obtained from the ground temperature network was extended in 1971 through the use of resistivity surveys (Barnes, 1966; D.K. MacKay, 1969). The Schlumberger configuration was used to determine the lower permafrost boundary, after successful checking at sites with ground temperature control data. Some profiling also was done using the Wenner array. Also, a limited amount of seismic work was carried out in 1970.

Major Features of Permafrost Distribution in the Study Area

Before proceeding with the application of heat conduction models, the major features of permafrost distribution are summarized. Where a permafrost "thickness" is quoted this is measured from the ground surface. Thus it includes the active layer, which is not correct by definition, but convenient. Where mean annual temperatures are quoted they refer to mean values calculated over the period of field observations (either one or two years). Temperature data were collected irregularly over this period.

Estimated from near surface measurements, the mean annual surface temperature for spruce-covered sites is in the order of -4.0 to -4.5°C; the mean annual river temperature (unfrozen portion) is close to +4.0°C, in lakes is about +3.2°C whilst for bare ground on slip-off slopes the mean annual surface temperature is about -1.0 to -1.5°C. Mean annual water temperatures are based on daily measurements for the period May to October, and an assumed winter value of 0.0°C (Table 2). In winter the unfrozen portions of the lakes are too shallow to permit any thermal stratification, and the water flowing in channels is mixed and cooled to a uniform temperature.

Spruce-Covered Areas

The greatest permafrost thicknesses are found in locations which have a mature spruce cover and are farthest away from any water bodies. Spruce-covered sites are areas of little geomorphic change. The maximum thickness of permafrost locally exceeds 65 m as indicated by temperature measurements in boreholes. Linear extrapolation of the temperature profile at site 6-3 (Fig. 4) yields a thickness of about 65 m; a seismic reflection was obtained at a depth of 66 m at this locality. Linear extrapolation is only strictly valid for the steady-state situation with homogeneous ground material. Table 3 shows thicknesses that have been determined based on extrapolation, for spruce-covered areas (for transect locations *see* Figs. 1, 2, and 3). In stable areas the

[1] Thermistors were calibrated after being sealed. The bridges used were calibrated regularly against a precision decade resistance. One source of error which cannot be accounted for, however, is that which could result from stress following installation in a borehole, as the cable freezes in place.

TABLE 3

Permafrost Thicknesses for Spruce-Covered Areas

Site number	Permafrost thickness (m)
2-1	52.0
2-8	50.0
6-2	65.0
6-3	65.0
6-17	53.0
SL-10	57.0

TABLE 4

Ground Temperatures Beneath a Cutbank and a Slip-off Slope

	Cutbank		Slip-off slope
Depth (m)	Site 6-1 (x = 1 m)	Site 6-2 (x = 36 m)	Site 6-5 (x = 35 m)
6	-2.8°C	-3.6°C	-0.5°C
9	-2.4	-3.3	0.0
12	-2.0	-3.0	+0.2
15	-1.7	-2.7	+0.4
20	-1.4	-2.4	+0.7

thermal effect of surface water bodies is exhibited as warmer mean annual ground temperatures in their vicinity, as shown in Table 4 (x is distance to channel).

Slip-Off Slopes

In locations indicative of contemporary geomorphic change, permafrost is much thinner and may even be entirely absent. Ground temperatures under the slip-off slope are warmer than those on the cutbank side of the river (Table 4). Table 5 shows permafrost thicknesses determined on slip-off slopes close to the channel. These sites are characterized by bare surfaces (commonly submerged in summer). With increasing distance from the river this gives way to a scattered growth of willow on the slip-off slope. This increase in vegetation leads to a concurrent lowering of ground temperatures and to the aggradation of permafrost (Tyrtikov, 1963; Viereck, 1970; Gill, 1971). The

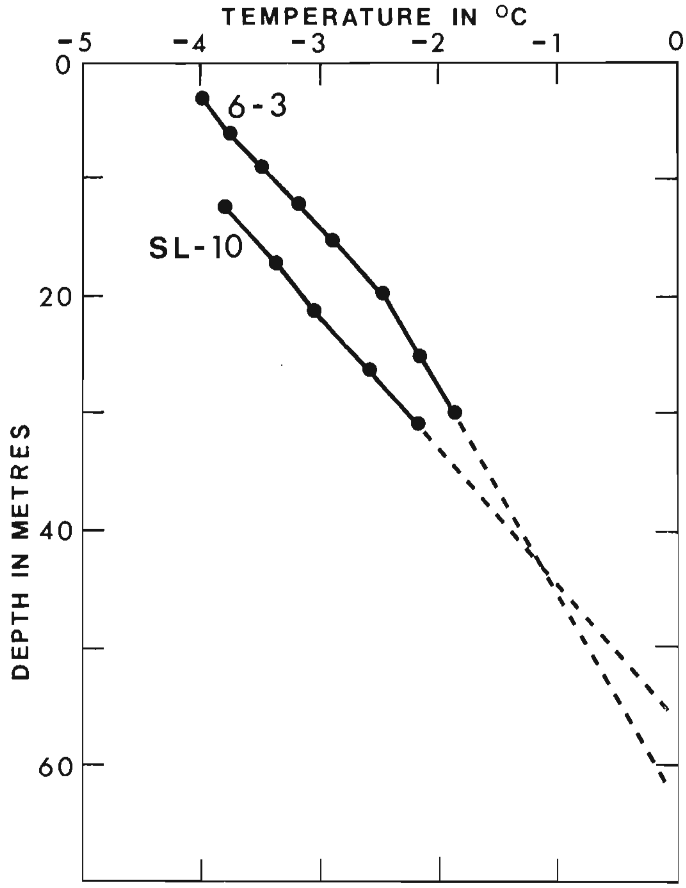

Figure 4. Temperature boreholes: spruce-covered sites.

pattern of permafrost aggradation however, is only partly caused by the insulating effect of the developing vegetation. For as the channel migrates laterally, a part of the surface which it previously occupied is exposed, and the ground there gradually will cool under the influence of mean annual surface temperatures below 0°C. As one moves across the slip-off slope the land surface is progressively older; consequently the degree of thermal recovery is more advanced and permafrost is thicker (Table 6).

The boreholes on the slip-off slope show that temperatures continue to increase with depth below the base of permafrost (Fig. 6), indicating that a through-talik possibly does exist beneath the channel. It is possible, though, that temperatures might decrease at even greater depths, so that only a pseudo-talik exists as suggested by Efimov (1964, p. 104).

Intermediate Areas

Between the slip-off slope and the spruce-covered area there is a surface of intermediate age dominated by a willow-alder association. Ground temperatures and permafrost thicknesses are intermediate also, with temperatures at 15-m depth ranging from -0.1°C (site 6-11) to -1.6°C (site 2-6) and permafrost thicknesses as shown in Table 7.

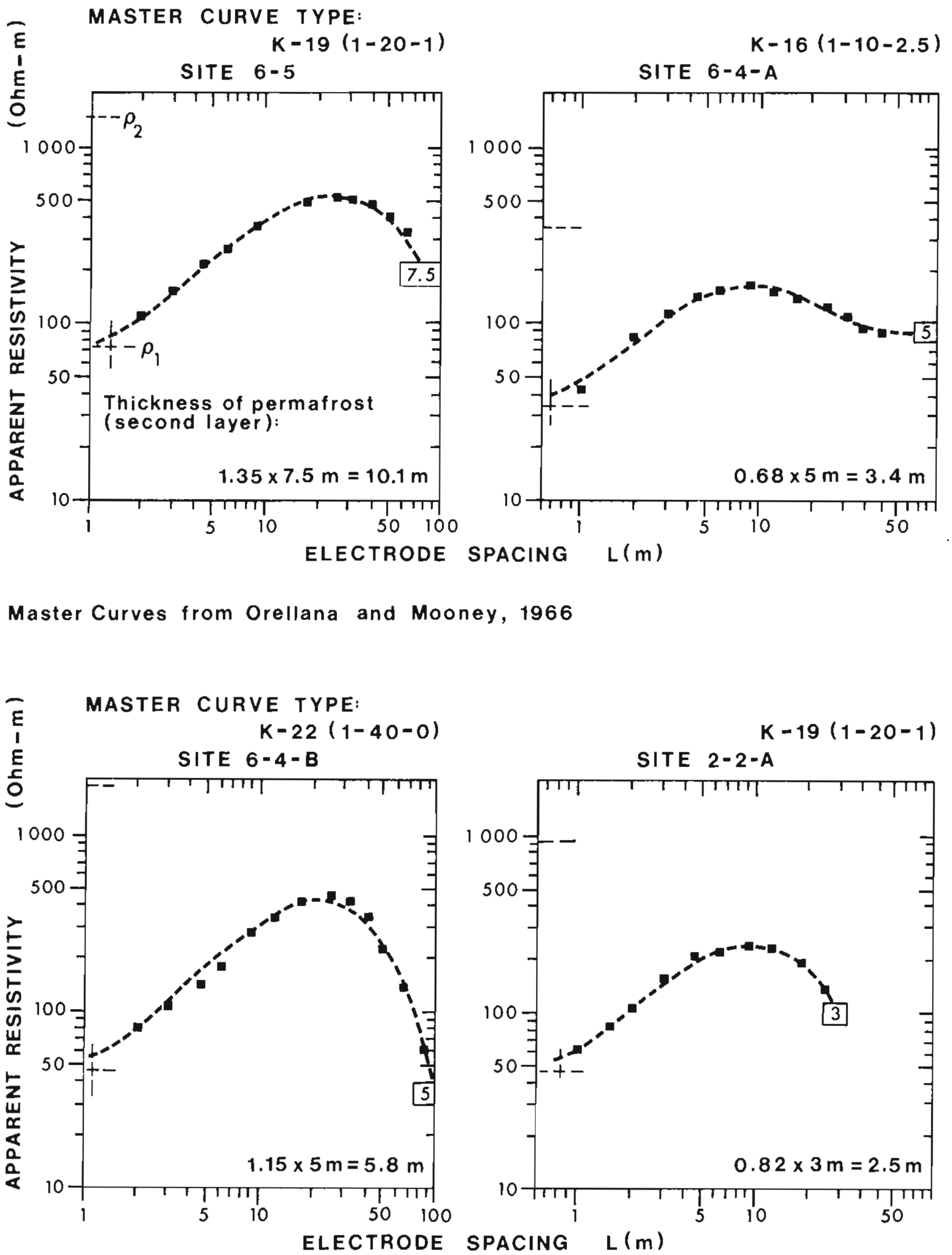

Figure 5. Resistivity sounding data for sites on slip-off slopes.

Snowbank Effects on Permafrost Distribution

The generalized pattern of permafrost aggradation outlined in Table 6 is actually interrupted by a zone of relatively temporary degradation (Fig. 7). The configuration of permafrost shown in Figure 7 was determined by probing (mostly confined to the top 3 m) and by temperature data. The former technique can be misleading at times; where temperatures in the frozen ground beneath the talik are very close to 0°C the ground will feel soft to the probe. Resistivity profiling was carried out across the slip-off slope using the Wenner array, and values of apparent resistivity for two ground layers, calculated by the Barnes layer method, are shown in Figure 8. At either end of the transect the values for both layers are representative of frozen ground (cf. D.K. MacKay, 1969, p. 372). For the upper layer (1 to 5 m) the values of less than 100

TABLE 5

Permafrost Thicknesses for Slip-off Slopes

Site number	Distance to channel (m)	Permafrost thickness (m)	
2-2-A	10	2.5	Resistivity sounding (*see* Fig. 5)
2-3	20	7.0	Temperature borehole
6-4-A	5	3.4	Resistivity sounding
6-4-B	15	5.8	Resistivity sounding
6-4	25	8.5	Temperature borehole
6-5	35	9.0	Temperature borehole

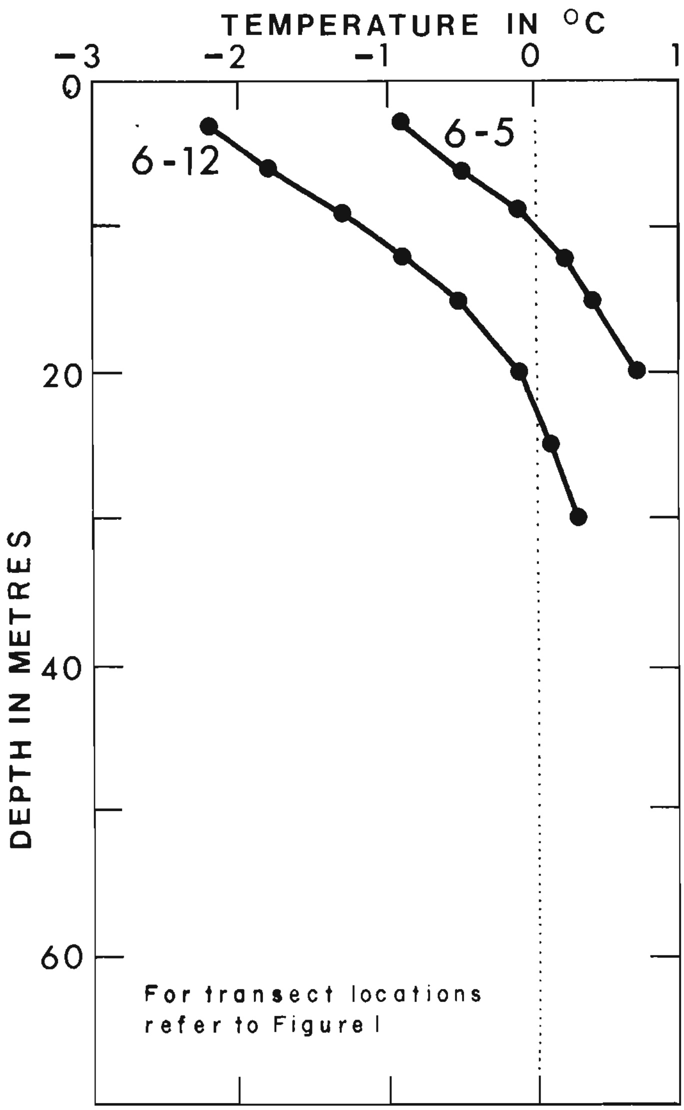

Figure 6. Temperature boreholes: slip-off slope sites.

ohm-metres in the middle of the transect are indicative of unfrozen ground. The values for the lower layer are higher, 270 to 540 ohm-metres, and could indicate the presence of a frozen layer, although close to 0°C. Dry, unfrozen ground also would produce high resistivity values.

The talik in Figure 7 coincides with the zone of maximum seasonal snow accumulation (cf. Gill, 1973). Along the slip-off slope the outer belt of willows (willow (1)) forms an excellent setting for the development of deep winter snow drifts. These are formed by the prevailing winds and the general shape is reproduced each year. As the willow colonizes new areas across the slip-off slope, the snowbank location migrates along with it. Ground temperatures in its lee progressively cool and permafrost wedges back--an analogous process to that which occurs in the case of river migration. The effect of snow cover on ground temperature is well documented (Shul'gin, 1957; Gold, 1958, 1963, 1967; Pearce and Gold, 1959) but detailed local studies of the influence of snow cover on permafrost are lacking. The temperatures in the permafrost under the slip-off slope are close to 0°C, which, combined with the gradual release of heat from the underlying sediments (heat which derives from the period when the river occupied this position), clearly makes it vulnerable to degradation. It is evident that mean annual surface temperatures in this snowbank zone are raised to near 0°C so that permafrost is "temporarily" degrading there. Thus the snow cover is important as a permafrost controlling factor here. This pattern is examined below in the context of heat conduction theory.

Summary

It is possible to recognize three concomitant sets of processes which affect the overall configuration of permafrost. Under steady-state conditions the thermal effect of water bodies depends upon the distance from

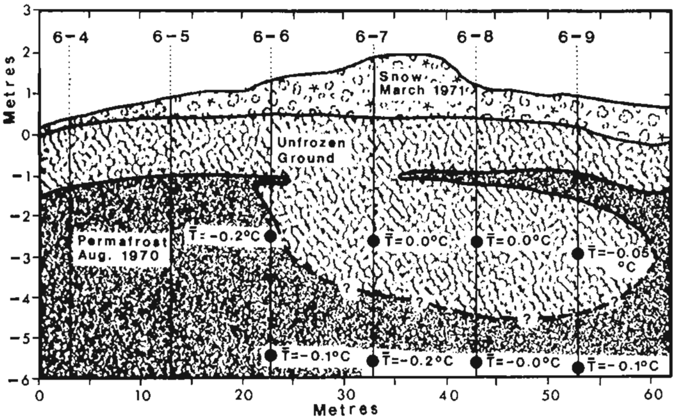

Figure 7. Details of permafrost configuration beneath a slip-off slope.

TABLE 6

Variation of Permafrost Thickness with Distance Away from River

Site[1] number	Distance from river (m)	Temperature at 9 m (°C)	Permafrost thickness (m)
6-4	25	+0.02	8.5
6-5	35	-0.02	9.0
6-7	55	-0.1	10.0
6-9	75	-0.1	12.0
6-11	100	-0.7	17.5
6-12	135	-1.2	22.5
6-13	170	-1.9	24.5

[1]See Figure 3 for borehole locations.

TABLE 7

Permafrost Thicknesses for Areas of Willow-Alder

Site number	Permafrost thickness (m)	Distance from back of slip-off slope (m)
2-5	29.0	3.0
2-6	38.0	22.0
6-11	17.5	2.0
6-12	22.5	37.0
6-13	24.5	72.0
6-14	29.0	107.0

them; in the case of the shifting channel (transient state) the distance relates to the degree of thermal recovery (i.e., the process of permafrost aggradation). Associated with this horizontal gradient of thermal recovery is a vegetation gradient which acts upon the ground thermal regime in the same sense. The major features of the resultant pattern of permafrost configuration are summarized in Figure 9. Lastly, superimposed on this configuration is the permafrost degradation associated with the snowbank.

GROUND TEMPERATURES AND HEAT CONDUCTION THEORY

Increased value can be obtained from detailed local thermal studies if it is possible to demonstrate that the ground temperature field displays some consistency with respect to environmental factors. This is now attempted through the framework of simple heat conduction theory.

Theory

If the mean surface temperature is the same everywhere and is steady with time, the mean temperature at any point in the ground is simply a function of the surface temperature $\bar{T}_s$ and the earth's geothermal gradient Gg. If $\bar{T}_s$ were below 0°C, permafrost would form and ultimately attain an equilibrium depth at which temperature increase due to internal earth heat just offsets the amount by which 0°C exceeds $\bar{T}_s$. (In practice the equilibrium configuration might not be approached for perhaps thousands of years, depending on the ultimate thickness.)

An important and intriguing problem, however, is to determine the disturbance of subsurface temperatures that results when the surface temperature within some finite region differs from that of the area outside the region. A solution to this problem for arbitrary-shaped regions (lakes, for example) has been derived by Lachenbruch (1957a):

$$\theta(x,y,z,t) = (\bar{T}_d - \bar{T}_s) \sum_{0}^{360} \frac{\lambda}{360} \left\{ \frac{1}{\sqrt{1+(R_1/Z)^2}} \operatorname{erfc}\left[\frac{\sqrt{Z^2+R_1^2}}{2\sqrt{\alpha t}}\right] - \frac{1}{\sqrt{1+(R_2/Z)^2}} \operatorname{erfc}\left[\frac{\sqrt{Z^2+R_2^2}}{2\sqrt{\alpha t}}\right] \right\} \quad (1)$$

where $\bar{T}_d$ is the mean surface temperature within the finite region. The temperature at any point in the ground is given by:

$$\bar{T}(x,y,z,t) = \theta(x,y,z,t) + \left\{ \bar{T}(x,y,0,t) + Gg.z \right\} \quad (2)$$

Here, the first term represents the sum of the temperature contributions at the common apex of sectors of a circular annulus, of central angle λ and inner and outer radii R_1 and R_2 (Fig. 10). The second term represents the normal (undisturbed) temperature profile for the area. For a point lying within the finite region, R_1

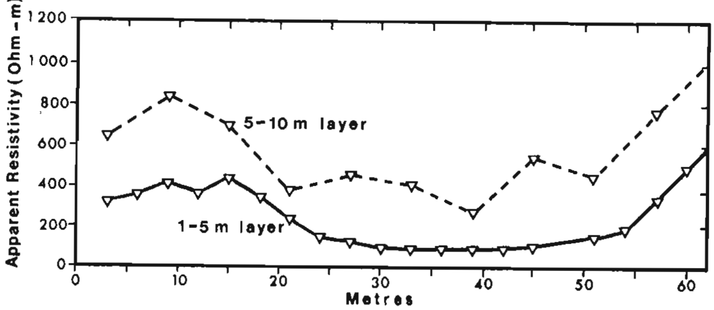

Figure 8. Resistivity profile across a slip-off slope.

For a homogeneous medium, and neglecting the effects of latent heat, an equation describing the thermal disturbance produced by a river whose temperature is $\bar{T}_d$, is (Lachenbruch, 1957b, p. 1517-1522):

$$\theta(x,z,t) = (\bar{T}_d-\bar{T}_s)\left\{\psi\left(\frac{x}{z},m\right) - \psi\left(\frac{x-s}{z},m\right)\right\} \qquad (4)$$

where the function

$$\psi\left(\frac{x}{z},m\right) = \frac{1}{2}\operatorname{erfc}(\sqrt{m}) + \frac{1}{\Pi}\int_{o}^{x/z} \frac{e^{-m(1+u^2)}}{1+u^2}\,du \qquad (5)$$

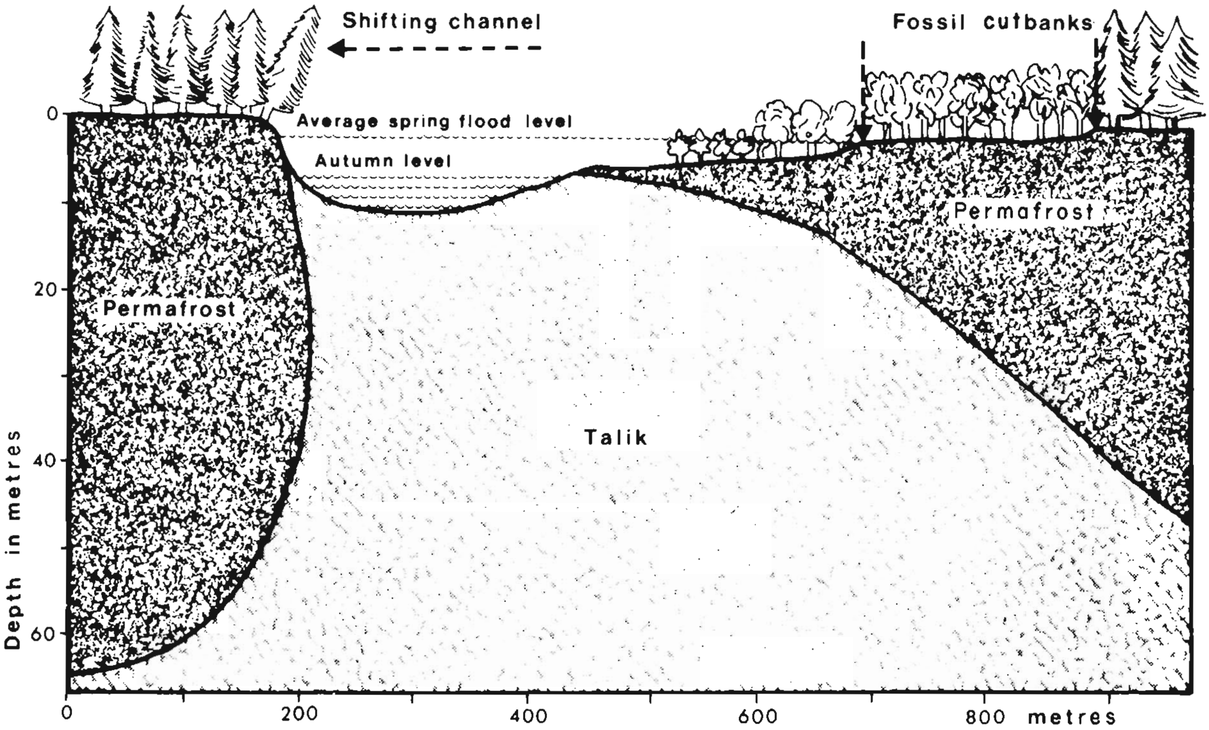

Figure 9. Cross-section through a shifting channel area.

is set to zero and R_2 is simply the radius of the body. For t = ∞ (steady state) equation (1) reduces to:

$$\theta(x,y,z) = (\bar{T}_d-\bar{T}_s)\sum_{o}^{360}\frac{\lambda}{360}\left\{\frac{1}{\sqrt{1+(R_1/Z)^2}} - \frac{1}{\sqrt{1+R_2/Z)^2}}\right\} \qquad (3)$$

This is the form used by W. G. Brown *et al.* (1964, p. 150). The basic solution has been employed in more elaborate form in a theory of pingo formation (Mackay, 1962).

Lachenbruch (1957b) presents an equation suitable for predicting the thermal effect of a strip-shaped disturbance such as a river. Assuming a steady-state condition, the configuration of taliks beneath rivers depends upon the heat balance of the river, the mean annual ground surface temperature, and the geothermal gradient. Whether or not the talik penetrates through the permafrost is also related to the width of the river. The time-dependent case involves consideration of α and t.

and where $m = z^2/(4\alpha t)$
x = horizontal distance from one side of the strip (river)
s = width of strip (river)
u = variable of integration.

Both $\bar{T}_d$ and $\bar{T}_s$ can be varied over discrete time intervals, thus making it possible to account, in some way, for climatic change, for example. For the steady-state, equation (5) reduces to:

$$\psi\left[\frac{x}{z},0\right] = \tfrac{1}{2} + \frac{1}{\pi}\int_{o}^{\frac{x}{z}} \frac{du}{1+u^2} = \tfrac{1}{2} + \frac{1}{\pi}\tan^{-1}\frac{x}{z} \qquad (6)$$

and equation (4) becomes:

$$\theta(x,z) = \frac{(\bar{T}_d-\bar{T}_s)}{\pi}\left\{\tan^{-1}\frac{x}{z} - \tan^{-1}\frac{x-s}{z}\right\} \qquad (7)$$

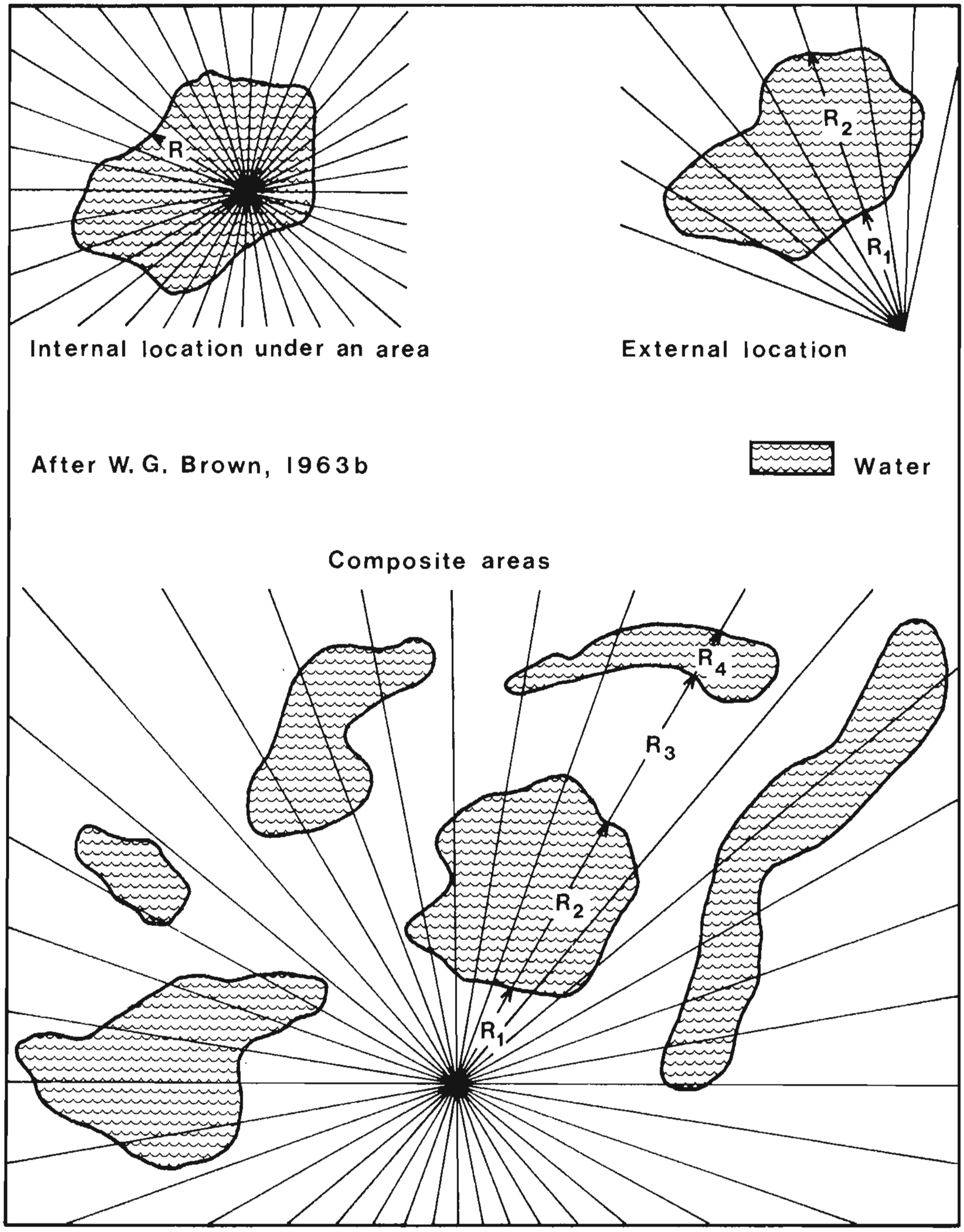

Figure 10. Method of dividing a given surface area into sectors of circles.

By way of illustration some calculations were made based on these equations using input data that are generally representative of conditions in the Mackenzie Delta. In the first group of examples (Fig. 11) computations were made for the steady-state solution (m = 0) for various values of s. The thermal disturbance was calculated from equation (7) and the ground temperature field from equation (2). It is evident that as the width of the river is increased, a critical width is reached beyond which a through-talik will exist beneath the river. In this steady-state case the critical width is about 70 m. As the width is increased further the proportion of unfrozen ground becomes progressively greater. While the upper surface of permafrost is lowered under the influence of the river, this is mirrored by a corresponding rise in the lower permafrost boundary.

In Figure 12 results are presented which show the recession of the permafrost table with time, beneath a river 100 m wide and assuming a thermal diffusivity of 0.060 m^2 per day. A higher diffusivity would advance the recession whereas a lower value would

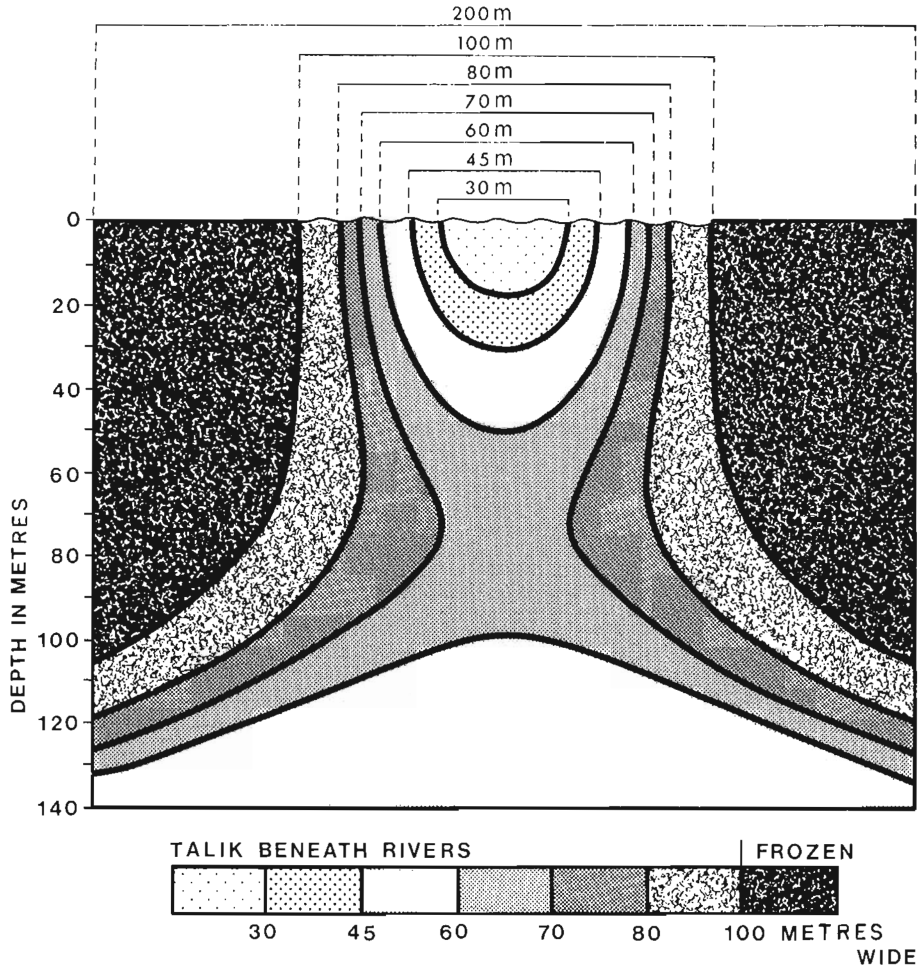

Figure 11.

Steady-state permafrost configuration under rivers.

slow it down. The computations presented in Figure 12 were made for a river at +4°C, which was assumed to have been instantaneously placed on ground with a surface temperature of -4°C at time-zero. The calculations used for Figures 11 and 12 reveal that in the vicinity of the 100-m-wide river, for depths up to 30 m for example, the steady-state configuration is virtually attained within 500 years (deviation of 0.1 to 0.2°C). For depths up to 60 m it would take 750 to 1000 years. A through-talik develops in 200 to 300 years. It is likely that through-taliks exist beneath channels which are more than 100 m wide.

Further calculations of the effect of lake no. 2 (*see* Fig. 1) on ground temperatures at site 6-3, at various times after the instantaneous introduction of the lake into the landscape, show that after 500 years the temperature effect is within, at most, 0.2°C of the steady-state solution. For depths up to about 20 m this difference is only 0.1°C or less (Table 8).

The models employed in this study assume conduction in a homogeneous medium whose thermal properties are not functions of temperature. To be consistent with this view of the problem it is considered more appropriate to use an apparent, overall value for α as input into the models, as opposed to the "true" values. Before this is done, however, the actual ground materials are described.

Ground Materials

As would be expected for a Delta environment, the grain-size distributions of the 50 samples tested fall within narrow limits. The material to a depth of 30 m consists mainly of silt and fine sand. The fact that most of the samples were 3-m integrated sections inevitably produced some homogenizing effect. Total moisture contents (Ww) on a dry-weight basis vary from almost 100 per cent in near surface layers to 30 per cent at depth. Total carbon content, as determined by loss on ignition, varied between only 5 per cent and 7 per cent for 30 samples. These data are similar to those of Johnston and Brown (1965). They also found that moisture content decreased with depth and that visible ice segregation was confined mostly to the upper 10 m. R.J.E. Brown (1956) reported that ground ice in some Delta soils at Aklavik was either cement ice or thin (up to 16 mm) veinlets. This was the pattern in the present study area with some ice wedges also being observed. Thermal properties of some of the field samples were calculated, assuming all the moisture to be ice and all the carbon to be organic matter (Table 9). Values for conductivity, k, were calculated using Kersten's (1949) formulae, and volumetric heat capacity, C, after de Vries (1963). The thermal diffusivity, α, is given by k/C.

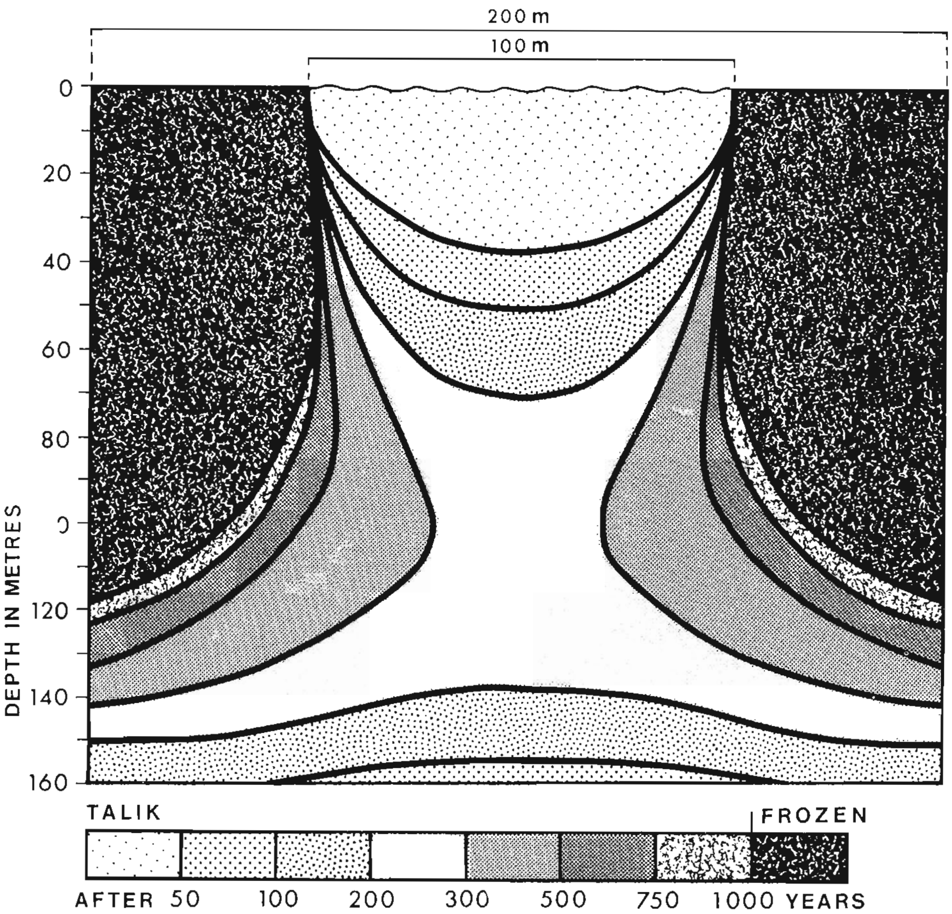

Figure 12. Permafrost regression, with time, under a river 100 m wide.

TABLE 8

Theoretical Thermal Effect (^{o}C) of Lake no. 2 on Ground Temperatures at Various Depths at Site 6-3 As a Function of Time[1]

Depth (m)	50	100	200	300	500	750	1000	Steady-state
5	0.08	0.12	0.14	0.15	0.17	0.17	0.18	0.20
10	0.16	0.23	0.28	0.31	0.33	0.35	0.36	0.40
15	0.23	0.33	0.40	0.44	0.47	0.50	0.51	0.60
20	0.27	0.40	0.50	0.54	0.60	0.64	0.66	0.74
25	0.30	0.45	0.58	0.64	0.70	0.74	0.77	0.87
30	0.31	0.48	0.64	0.70	0.78	0.83	0.86	0.99

[1]Thermal diffusivity, α, was taken as 0.06 m^2 per day, and ($\bar{T}_d - \bar{T}_s$) was 7.4°C; site 6-3 is 50 m from lake no. 2.

TABLE 9

Physical and Thermal Properties of Some Soil Samples

Borehole	Depth interval (m)	Soil type*	Ww%	Xo%	B** ($g\ cm^{-3}$)	k ($W\ cm^{-1}\ C^{-1}$)	C ($J\ cm^{-3}$)	α ($cm^2\ sec^{-1}$)	α ($m^2\ day^{-1}$)
SL-1	0-3	silt	63.3	(6.0)	1.35	0.0372	2.0005	0.0188	0.1625
	3-6	silt	46.9	(6.0)	1.38	0.0292	2.1081	0.0139	0.1197
	6-9	silt	30.0	(6.0)	1.42	0.0200	2.2227	0.0094	0.0810
	9-12	sandy silt	34.1	(6.0)	1.45	0.0235	2.1872	0.0108	0.0930
	12-15	silt	35.9	(6.0)	1.49	0.0260	2.1662	0.0111	0.0959
	15-18	silt	36.7	(6.0)	1.52	0.0238	2.1544	0.0128	0.1106
SL-9	0-3	silt	39.4	6.0	1.35	0.0238	2.1670	0.0110	0.0949
	3-6	silt/sand	37.7	6.5	1.38	0.0237	2.1713	0.0109	0.0943
	6-9	sandy silt	39.8	6.0	1.42	0.0262	2.1505	0.0122	0.1054
	9-12	clayey silt	33.3	5.7	1.45	0.0230	2.1949	0.0105	0.0906
	12-15	clayey silt	33.7	5.6	1.49	0.0244	2.1853	0.0112	0.0969
SL-10	0-3	clayey silt	90.0	(6.3)	1.00	0.0352	1.9766	0.0178	0.1536
	3-6	silt	40.0	6.3	1.38	0.0251	2.1555	0.0116	0.1005
	6-9	clayey silt	34.0	(6.3)	1.42	0.0226	2.1916	0.0103	0.0891
	9-12	sandy silt	30.5	6.3	1.45	0.0212	2.2125	0.0096	0.0828
	12-15	sandy silt	32.2	(5.5)	1.49	0.0213	2.1976	0.0107	0.0923
	15-18	silty sand	30.6	4.9	1.52	0.0235	2.2087	0.0105	0.0911
	18-21	silty sand	31.8	(5.0)	1.55	0.0251	2.1934	0.0114	0.0988
	21-24	silty sand	33.0	5.4	1.57	0.0267	2.1780	0.0122	0.1057
	24-27	sandy silt	33.9	(5.5)	1.59	0.0292	2.1665	0.0129	0.1118
	27-30	sandy clayey silt	31.0	6.1	1.61	0.0265	2.1843	0.0121	0.1049

* Grain-size classes are per U.S.D.A. classification. Where only a single class appears this indicates that neither of the other two exceeds 20%. When a class(es) exceeds 20%, but is not the main one, it appears as the modifier.

** Values supplied by R.J.E. Brown (pers. comm.). All thermal properties pertain to the frozen state.

Apparent Thermal Diffusivity

If an appropriate ground temperature record is available (observations of T vs. z for different times), it is possible to analyze it, consistent with the assumptions of heat conduction theory, and to compute the value of α for which the observations best fit the theory. It is this apparent value which is used in subsequent models.

Methods based on the periodic flow of heat in a homogeneous medium have been used to infer certain information on ground thermal properties from temperature records (Ingersoll *et al.*, 1954; Carson, 1963; van Wijk, 1963). The method used here is based on that of Lovering and Goode (1963), which itself is based on the intersection points of two tautochrones. The vertical separation of successive crossings of any two curves is always equal to one-half the wavelength (Fig. 13); if this can be measured the diffusivity can be easily calculated. In practice the small range of actual temperatures at the depths of the second crossing makes its precise determination very difficult (Fig. 13). An alternative procedure is presented by Lovering and Goode when two or more temperature-depth curves are available and the times t_1 and t_2 (after t = 0) at which they were measured are known. It is thus mandatory to know the date for t = 0, i.e., the time in spring when the annual surface temperature wave passes through the mean value. Thermal diffusivity is then calculated from:

$$\alpha = \frac{4x_c^2\ \pi/P}{\left[2(t_1\ \pi/P + t_2\ \pi/P) \pm n\pi\right]^2} \quad m^2\ day^{-1} \qquad (8)$$

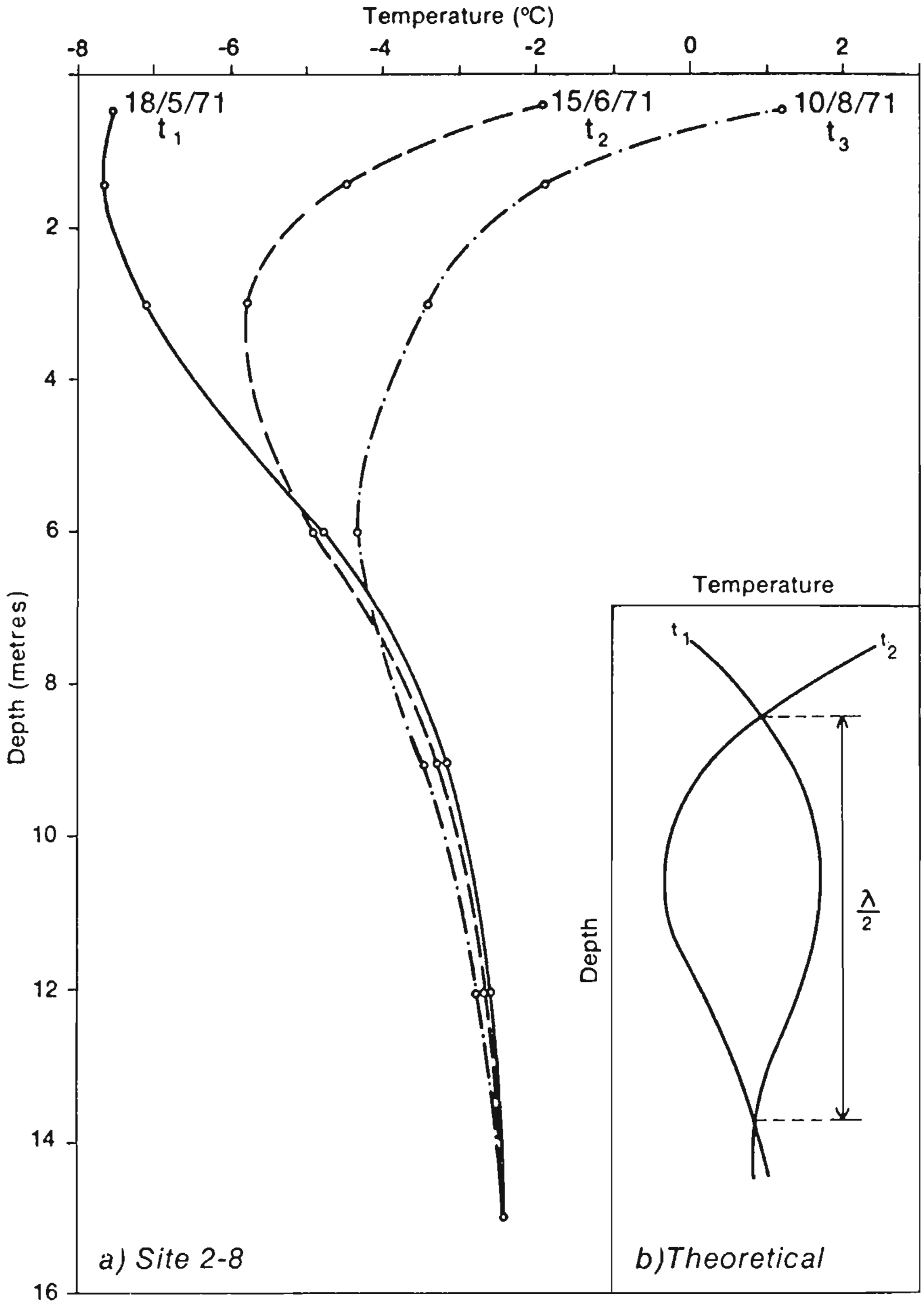

Figure 13. Examples of temperature-depth curves used for calculation of apparent thermal diffusivity.

where x_c is the depth of the first crossing point, P the period of the wave, and n is zero or any integer (Lovering and Goode, 1963, p. 42). Since the date of t = 0 was not accurately known in the present study, a variation of the method was devised. Given equation (8) it can be shown that when three tautochrones (t_1, t_2, t_3) are available,

$$\alpha = \frac{(x_c' - x_c)^2}{(t_3 - t_2)^2} \cdot \frac{P}{\pi} \qquad m^2 \quad day^{-1} \qquad (9)$$

TABLE 10

Calculations for Apparent Diffusivity
Using Data From Boreholes 2-6 and 2-8

Tautochrone dates	Pair number	x_c(m) no. 2-6	x_c(m) no. 2-8
18/5/70 + 4/6/70	1	4.90	5.30
18/5/70 + 15/6/70	2	5.15	5.55
18/5/70 + 12/7/70	3	5.85	6.25
18/5/70 + 10/8/70	4	6.55	6.85
15/6/70 + 12/7/70	5	6.50	6.80
15/6/70 + 10/8/70	6	7.10	7.40

"$t_3 - t_2$"			no. 2-6		no. 2-8	
Tautochrone dates	Days	By combining pairs no.	$(x'_c - x_c)$	α m^2 day^{-1}	$(x'_c - x_c)$	α m^2 day^{-1}
15/6 – 4/6	11	(2) – (1)	0.25	0.060	0.25	0.060
12/7 – 4/6	38	(3) – (1)	0.95	0.073	0.95	0.073
10/8 – 4/6	67	(4) – (1)	1.65	0.070	1.55	0.062
12/7 – 15/6	27	(3) – (2)	0.70	0.078	0.70	0.078
10/8 – 15/6	56	(4) – (2)	1.40	0.073	1.30	0.062
10/8 – 12/7	29	(4) – (3)	0.70	0.067	0.60	0.050
12/7 – 18/5	55	(5) – (2)	1.35	0.070	1.25	0.060
15/6 – 18/5	28	(5) – (3)	0.65	0.062	0.55	0.045
10/8 – 18/5	84	(6) – (2)	1.95	0.062	1.85	0.056
10/8 – 12/7	29	(6) – (5)	0.60	0.050	0.60	0.050
				$\bar{\alpha}$ = 0.067 s.d. = ±0.008		α = 0.060 s.d. = ±0.009

where x_c is the crossing point of one pair and x'_c the crossing point of the other. Calculations were carried out on data from two boreholes in different terrain segments, and results are presented in Table 10. The period P was taken as 365 days. Only where crossing points could be confidently located were the values used; where tautochrones happen to cross below about 7 m, they cross so obliquely that it is generally very difficult to fix the crossing point accurately. The overall mean value of 0.064 m^2 per day compares closely to the value of 0.067 m^2 per day given by W.G. Brown *et al.* (1964) for the Mackenzie Delta.

The values of α determined in this manner are lower than those in Table 9. A possible explanation for this is that latent heat effects in the active layer tend to slow down the penetration of isotherms, either during thawing or freezing. Since the above method does not treat these effects explicitly, the reduced rate of penetration is expressed in terms of a lower diffusivity.

PRESENT APPLICATION OF HEAT CONDUCTION THEORY

Steady-State Solution

This is the simplest application of the heat conduction model, involving only the geometry of the problem. Use of this solution (equation (3)) assumes that mean annual ground surface temperatures have not changed over time and that the distribution of land and water likewise has not changed. Sample calculations have shown that, for depths up to 30 m, the steady-state configuration is very closely approached within 500 years. Even for this period of time, though, the above assumptions cannot be considered altogether valid. Unfortunately there is no basis for any quantitative expression of possible long-term climatic variations in this region. Some Mackenzie Delta tree ring data (abstracted from Giddings, 1947) were analyzed.

TABLE 11

Observed (T_{obs}) and Predicted (T_{pre}) Temperatures - Cutbank Section

		TEMPERATURE CONTRIBUTION (°C)					
Site	Depth (m)	Lakes	Rivers	Gg	Total	T_{pre}	T_{obs}
6-1*	6	0.07	1.22	0.15	1.44	-2.8	-2.8
	9	0.10	1.64	0.23	1.97	-2.2	-2.4
	12	0.14	1.95	0.30	2.39	-1.8	-2.0
	15	0.17	2.16	0.38	2.71	-1.5	-1.7
	20	0.24	2.38	0.50	3.12	-1.1	-1.4
	25	0.29	2.49	0.63	3.43	-0.8	-
	30	0.34	2.57	0.75	3.66	-0.5	-
	40	0.44	2.59	1.00	4.03	-0.2	-
	50	0.54	2.54	1.25	4.33	+0.1	-
6.2*	6	0.11	0.25	0.15	0.51	-3.7	-3.6
	9	0.16	0.36	0.23	0.75	-3.4	-3.3
	12	0.23	0.48	0.30	1.01	-3.2	-3.0
	15	0.29	0.61	0.38	1.28	-2.9	-2.7
	20	0.36	0.76	0.50	1.62	-2.6	-2.4
	25	0.45	0.91	0.63	1.99	-2.2	-2.1
	30	0.52	1.05	0.75	2.32	-1.9	-1.8
	40	0.67	1.26	1.00	2.93	-1.3	-
	50	0.79	1.40	1.25	3.44	-0.8	-
	60	0.90	1.51	1.50	3.91	-0.3	-
	70	1.00	1.61	1.75	4.36	+0.2	-
6.3*	6	0.25	0.12	0.15	0.52	-3.7	-3.8
	9	0.37	0.17	0.23	0.77	-3.4	-3.5
	12	0.48	0.23	0.30	0.01	-3.2	-3.2
	15	0.67	0.29	0.38	1.34	-2.9	-2.9
	20	0.75	0.39	0.50	1.64	-2.6	-2.5
	25	0.89	0.47	0.63	1.99	-2.2	-2.2
	30	1.01	0.55	0.75	2.31	-1.9	-1.9
	40	1.19	0.71	1.00	2.90	-1.3	-
	50	1.31	0.84	1.25	3.40	-0.8	-
	60	1.39	0.95	1.50	3.84	-0.3	-
	70	1.46	1.03	1.75	4.24	0.0	-

* Mean annual surface temperature = -4.2°C

For lakes, therefore, $(T_d - T_s)$ = 7.4°C

For rivers, $(T_d - T_s)$ = 8.2°C

Values for ring widths were taken from his Figure 1 and a regression analysis against time was performed. Without implying any specific relationships to climate, no significant linear trend in tree growth rates between 1460 and 1940 was detected. The second assumption is not valid in the case of the shifting river channel where geomorphological change obviously is taking place. Also, most lakes in the area show some evidence of morphological change either by sedimentation, wave action, or thermal erosion. A simple model, however, can provide a useful first approximation to the ground temperature field, with the advantage that it requires little input data.

Through the use of equation (3) the total thermal effect, θ, of all the water bodies in the area on the ground temperatures at any point and at any depth, can be assessed (cf. W.G. Brown *et al.*, 1964). The predicted temperature for any (x, y, z) then can be determined from equation (2). Lake and river outlines were converted into digital (co-ordinate) form and a computer program written to calculate the θ term. The base map was constructed from the 1967 air photograph

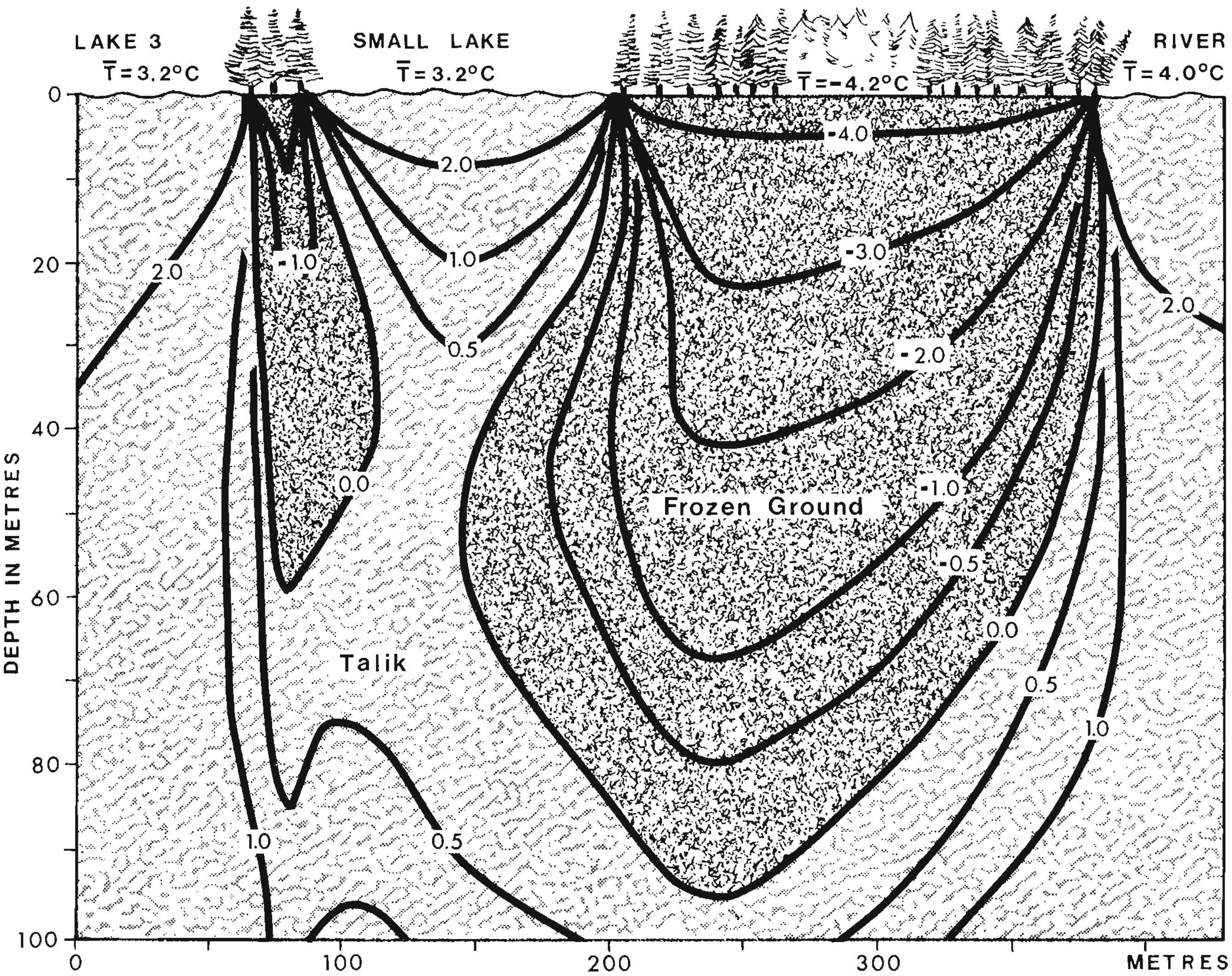

Figure 14. Computed temperature field under a traverse line.

which was taken during the spring flood. Outlines of water bodies then were reduced, on the basis of field data, by excluding areas not submerged during the rest of the year. (Use of the outlines straight off the air photograph would lead to an over-estimation of the θ term.) The computer program accommodates outlines of completely arbitrary shape, and the specified point (x, y, z) can be located anywhere, either inside or outside a water body. Distinction was maintained between lakes and rivers; from field data, values of 3.2°C and 4.0°C were taken to be representative of lake and river mean annual temperatures. Small lakes known to freeze through in winter were arbitrarily assigned a value of 0°C. Ground surface temperatures were estimated from near-surface measurements.

A value for the geothermal gradient, Gg, was calculated from the data of W.G. Brown *et al.* (1964) who worked in the Delta 50 km southeast of the present study area. They computed the thermal disturbance θ for an (x, y, z) and by knowing all the other parameters, solved equation (2) for Gg. However, they assumed a water temperature, $\bar{T}_W$, of 33°F (0.6°C) which is probably too low; this yielded a rather high value of 3.4°C/100 m for Gg. Their data have been reworked in the present paper using 3.6°C for $\bar{T}_W$; this gives an average value of 2.5°C/100 m for Gg. An average temperature gradient of about 2.3°C/100 m has been measured in a deep borehole some 35 km northwest of the study area (Jessop, 1970). Although the two sites are certainly different, the deep borehole result is taken as confirmation of the general validity of the 2.5°C value.

The model was first applied to boreholes 6-1, 6-2, and 6-3 on the cutbank side of the river. A value of -4.2°C for the surface temperature, $\bar{T}_S$, for spruce-covered areas was used. Such sites are the most stable and thus should be the best suited to the model. Observed and predicted temperatures are presented in Table 11, and agreement is very close. The slightly higher predicted values for site 6-1 are at least partly due to an over-estimate of the river effect; the river, of course, has not been steady in its present position

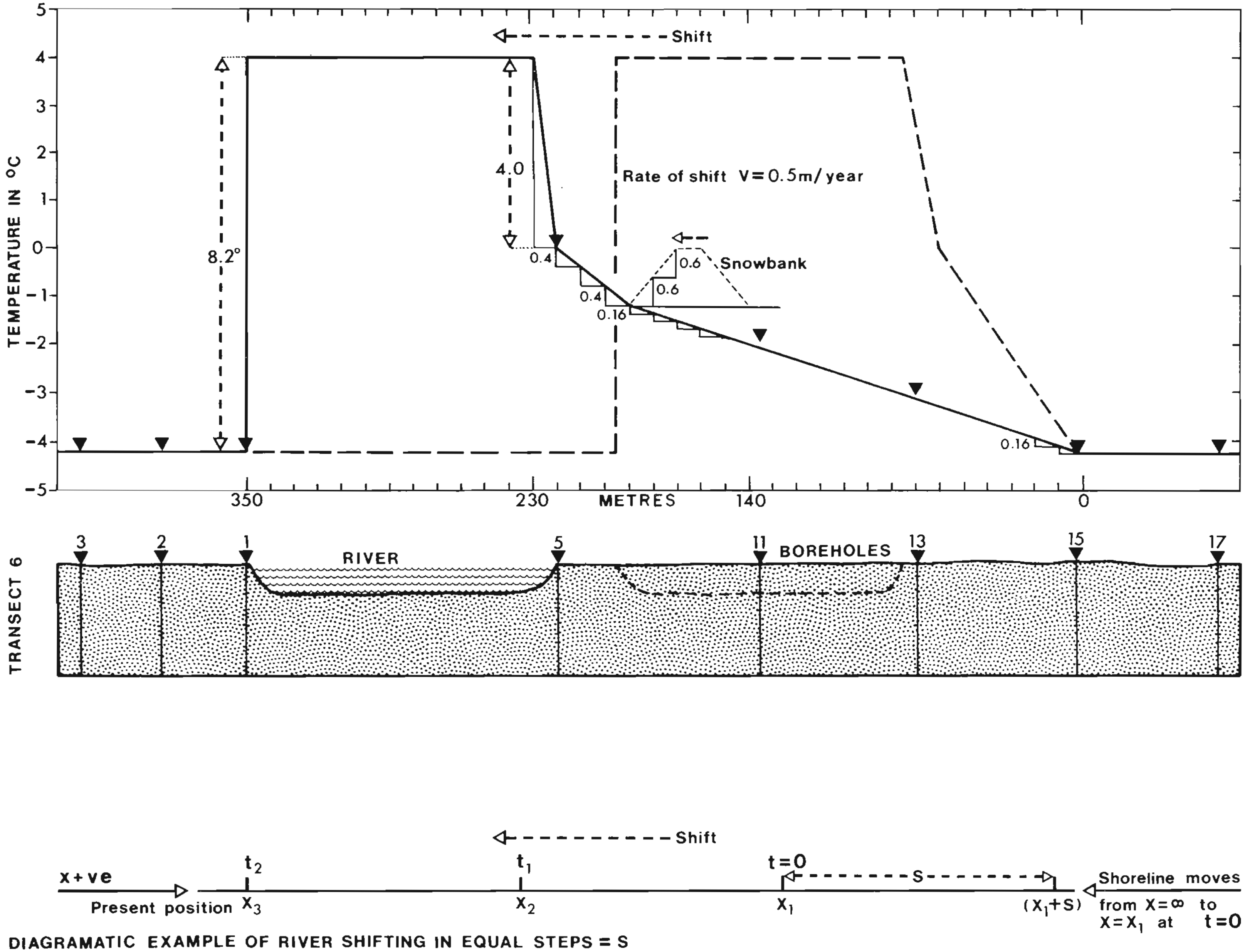

Figure 15. Temperature wave simulating river shifting.

TABLE 12

Observed and Predicted Temperatures (^{o}C) for the River-shifting Problem

DEPTH (m)	SITE 6-1		SITE 6-2		SITE 6-5		SITE 6-7		SITE 6-9		SITE 6-11		SITE 6-12	
	Obs	Pre	Obs	Pre	Obs	Pre	Obs	Pre	Obs	Pre	Obs	Pre	Obs	Pre
5	-3. 0	-2. 7	-3. 7	-3. 8	-0. 6	-0. 5	-0. 1	-0. 1	-0. 1	-0. 6	-1. 5	-1. 7	-2. 0	-2. 3
10	-2. 2	-1. 9	-3. 2	-3. 4	0. 0	-0. 2	0. 0	0. 0	-0. 1	-0. 5	-0. 7	-1. 4	-1. 1	-2. 0
15	-1. 7	-1. 5	-2. 7	-3. 1	0. 4	0. 1	0. 2	0. 1	0. 1	-0. 4	-0. 1	-1. 1	-0. 5	-1. 7
20	-1. 4	-1. 3	-2. 4	-2. 7	0. 7	0. 4	0. 4	0. 2	0. 2	-0. 2	0. 1	-0. 8	-0. 1	-1. 5
25	-	-1. 1	-2. 1	-2. 4	-	0. 5	-	0. 3	-	-0. 1	-	-0. 5	0. 1	-1. 2
30	-	-1. 0	-1. 8	-2. 2	-	0. 7	-	0. 5	-	0. 1	-	-0. 3	0. 3	-1. 0

for a long period of time. For each of these boreholes the total thermal contribution of all water bodies is generally of the order of one-and-one-half to two times as great as the earth's geothermal gradient. The thickness of permafrost in the vicinity of boreholes 6-2 and 6-3 is estimated at about 65 m, which correlates well with a seismic reflection at 66 m for this locality. Under the influence of the earth's geothermal gradient alone the 30-m temperature at borehole 6-3 would be about -3. 4^{o}C, and permafrost would be about 170 m thick.

On the basis of the close agreement above, mean temperatures were predicted beneath a traverse through a stable area for which no temperatures were available. Having once established values for the mean annual lake, river, and ground surface temperatures and the geothermal gradient, the entire ground temperature field can be calculated using equations (2) and (3). A value of -4. 2^{o}C was used for the ground surface temperature, 3. 2^{o}C and 4. 0^{o}C for the lake and river temperatures respectively. The calculated annual isotherms are shown in Figure 14.

The maximum permafrost thickness along the traverse is 95 m; the maximum thickness in the vicinity of the traverse is 102 m. These values are in excellent agreement with the data of Johnston and Brown (1964). Permafrost beneath the isthmus reaches only to 60 m. The permafrost shows the characteristically steep plunge at the edge of the river and lake no. 3. Under the small lake, however, although the boundary plunges initially, the thermal effect here is great enough only to form an "hour-glass shaped" talik. The upper permafrost surface is much depressed, while the lower one is raised (*see* also the 0. 5^{o}C isotherm), with only a narrow chimney actually penetrating the permafrost. If lake no. 3 were not so close there would be no through-talik beneath the small lake; the temperature at 50 and 60 m is only 0. 1^{o}C. Hand borings to depths of 1. 8 to 3. 4 m, made from the ice surface in a small lake nearby (lake no. 4) in February 1968 (C. P. Lewis and D. Gill, pers. comm.), indicate that the permafrost boundary plunges steeply at the shoreline.

As lake size increases, a critical radius will be reached beyond which no permafrost will be present beneath the lake centre. This critical radius, where permafrost opens up, can be calculated for steady-state conditions following the method of Mackay (1963, p. 79-80). Calculations have been made for various values of the geothermal gradient, Gg, for a typical case in the Mackenzie Delta. With Gg equal to 2. 5^{o}C/100 m the critical diameter is 166 m. This would be the case for a single lake with a temperature of 3. 2^{o}C in the midst of a region where the surface temperature was -4. 2^{o}C. As mentioned above, however, the combined thermal effect of many lakes in the Delta is to effectively increase the value of Gg. Therefore, additional calculations were made for the values of Gg of 3. 75^{o}C/100 m and 5^{o}C/100 m; the critical lake diameter for these cases turns out to be 111 m and 83 m. The critical width of a river whose temperature is 4. 0^{o}C is about 103 m and 77 m for these two cases. In the Mackenzie Delta, therefore, there should be an unfrozen chimney beneath lakes larger than 80 to 100 m in diameter (cf. Fig. 14) and an unfrozen curtain beneath rivers wider than about 80 m.

When this simple model was applied to sites on slip-off slopes, agreement was less satisfactory. In all cases the predicted temperatures were too cold, and permafrost thicknesses thus were over-estimated. At site 2-6, for example, the observed temperature gradient is 0. 12^{o}C/m, whereas the predicted value is only 0. 06^{o}C/m. A major thermal contribution thus is being underestimated by the model. Agreement is best close to the river; for example, at site 6-5 (Fig. 3), the observed and predicted 20-m temperatures are 0. 7^{o}C and 0. 0^{o}C. Farther from the river the results become poorer; at site 6-12 the observed and predicted 30-m temperatures are 0. 3^{o} and -1. 5^{o}C and at

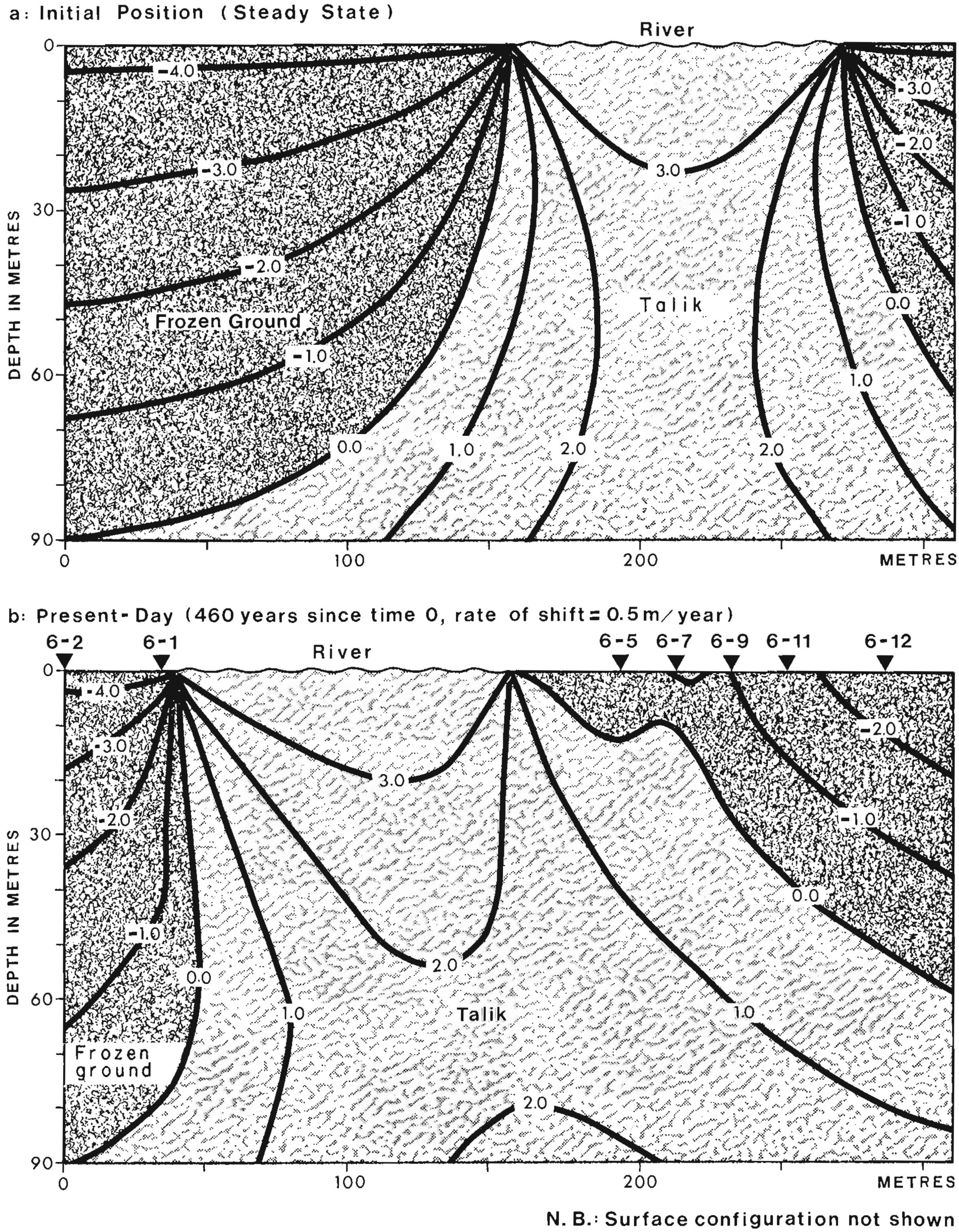

Figure 16. Permafrost history under a shifting channel.

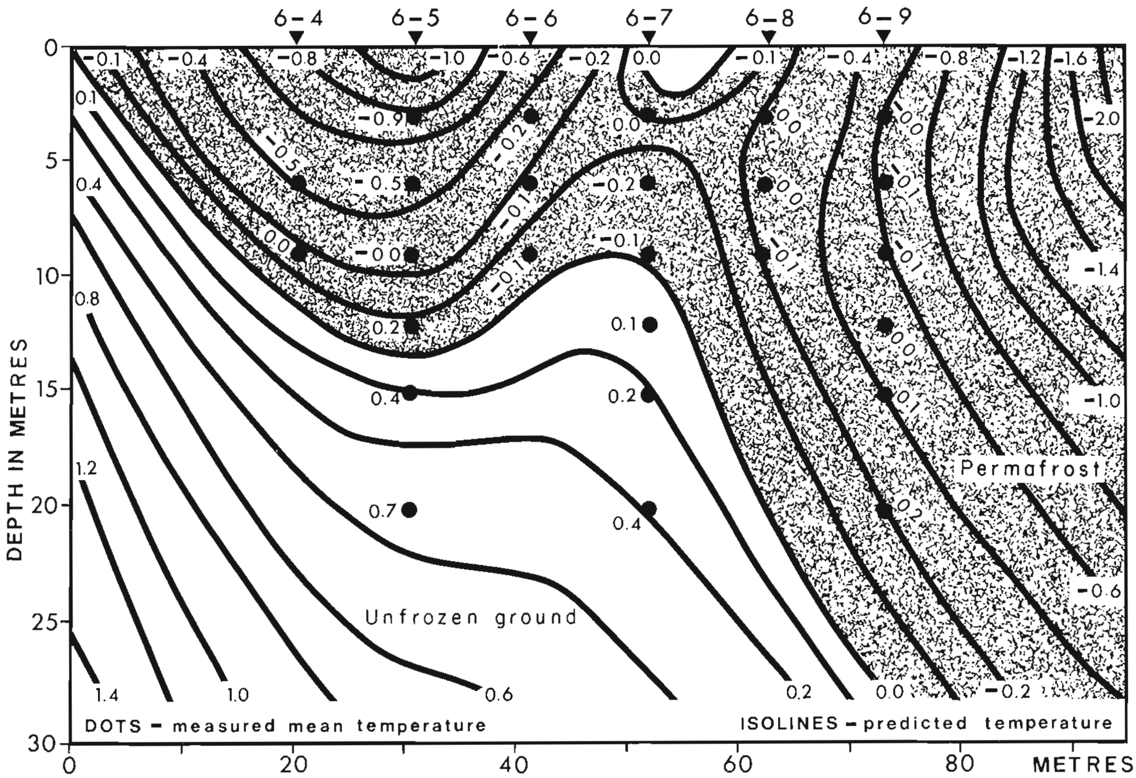

Figure 17. Predicted mean annual ground temperatures in the vicinity of the snowbank zone, slip-off slope.

site 6-14 they are 0.0°C and -2.3°C. This model is based on the assumption that the present conditions have existed unchanged for a long period of time, whereas sites in the lee of river migration have become progressively colder with time. Thus, at site 6-12, T_s has not always been as cold as the -2.6°C temperature used here, but in the recent past it was warmer (*see* Fig. 15). The ongoing cooling trend is illustrated in the shape of temperature-depth curves (cf. Figs. 4 and 6).

Transient State Solution

The steady-state model underestimates the river's thermal effect on the slip-off slope, and thus incorporation of the transient term associated with river migration is necessary. In reality river migration is a continuous process, but an attempt is now made to approach the real situation by treating the migration as a series of small, finite steps.

From the viewpoint of thermal history the migration of the channel can be simulated as a temperature-wave disturbance travelling across the surface (Fig. 15). This wave is not symmetrical, however, since the aggrading surface in the rear of the wave is subject to a sequential development of surface conditions, i.e., the succession of vegetation. The shape and amplitude of the wave was formed according to measured ground temperature data, as shown in Figure 15. The wave is moved across the surface from some initial position to its present position at a rate v, thought to approximate the average rate of shift over a long period of time (400 to 500 years). The initial position was taken as the older of two fossil cutbanks (Fig. 3); between this and the present position is another cutbank, indicating that the shifting has not been uniform. There has been an average rate of shifting of about 1.1 m per year over the past 30 years. It was decided, however, to use a lower "apparent" rate of shift in the model; this would incorporate some unknown lag time involved in the formation of the second cutbank by representing the shifting process as some equivalent uniform rate. If the river is assumed to have been shifting at a uniform rate of 0.5 m per year, the time between the initial and present positions would be 460 years. The thermal history, for example, could be computed on the basis of 10-m steps, i.e., corresponding to 20-year intervals.

In order to reproduce the thermal history mathematically, assumptions have to be made regarding the thermal regime which existed when the channel was in its initial position. The initial condition for this model was taken to be the steady-state solution with the channel in its initial position prior to migration.

Lachenbruch (1957b, p. 1521) gives an equation for θ in the case of an ocean which has transgressed or regressed by a number of stages of movement. For the shifting river (*see* Fig. 15), if three stages (t_0 (=0), t_1, t_2) are involved for example, the equation would be:

TABLE 13

Observed and Predicted Temperatures with Transient Correction for Boreholes on a Slip-off Slope

Temperature Contribution (^{o}C)

Site	Depth (m)	Gg(1)	Shifting channel* (i)	(ii)	(iii)	Total(2)	Other (3) rivers, lakes	Total (1)+(2)+(3)	T_{pre}	T_{obs}	Shifting**(4) channel
2-5	3	0.08	+2.15	-1.31	+0.01	+0.85	0.04	0.97	-2.0	-1.8	0.06
($\bar{T}_s$ = -3.0^{o}C)	6	0.15	+2.71	-1.59	+0.02	+1.14	0.09	1.38	-1.6	-1.6	0.12
	9	0.23	+2.90	-1.65	+0.03	+1.28	0.13	1.64	-1.4	-1.3	0.18
	12	0.30	+2.98	-1.64	+0.04	+1.38	0.18	1.86	-1.1	-1.0	0.24
	15	0.38	+3.01	-1.59	+0.05	+1.47	0.23	2.08	-0.9	-0.8	0.30
	20	0.50	+3.02	-1.49	+0.06	+1.59	0.31	2.40	-0.6	-	0.40
	30	0.75	+2.96	-1.26	+0.08	+1.78	0.45	2.98	0.0	-	0.57
2-6	3	0.08	+0.30	-0.13	+0.00	+0.17	0.05	0.30	-2.7	-2.8	0.05
($\bar{T}_s$ = -3.0^{o}C)	6	0.15	+0.58	-0.26	+0.01	+0.33	0.11	0.59	-2.4	-2.5	0.09
	9	0.23	+0.84	-0.37	+0.01	+0.48	0.17	0.88	-2.1	-2.2	0.14
	12	0.30	+1.06	-0.46	+0.02	+0.62	0.22	0.14	-1.9	-1.9	0.18
	15	0.38	+1.26	-0.52	+0.02	+0.76	0.29	1.43	-1.6	-1.6	0.23
	20	0.50	+1.51	-0.60	+0.03	+0.94	0.37	1.81	-1.2	-	0.30
	30	0.75	+1.84	-0.64	+0.04	+1.24	0.55	2.54	-0.5	-	0.43
	40	1.00	+2.00	-0.59	+0.04	+1.45	0.70	3.15	0.2	-	0.56
2-8	3	0.08	+0.06	-0.00	0.00	+0.06	0.17	0.31	-3.9	-4.0	0.02
($\bar{T}_s$ = -4.2^{o}C)	6	0.15	+0.11	-0.01	0.00	+0.10	0.34	0.59	-3.6	-3.7	0.05
	9	0.23	+0.16	-0.01	0.00	+0.15	0.51	0.89	-3.3	-3.3	0.07
	12	0.30	+0.22	-0.01	0.00	+0.21	0.66	1.17	-3.0	-2.8	0.09
	15	0.38	+0.27	-0.02	0.00	+0.25	0.83	1.46	-2.7	-2.5	0.12
	20	0.50	+0.35	-0.02	0.00	+0.33	1.05	1.88	-2.3	-	0.16
	30	0.75	+0.51	-0.03	0.00	+0.48	1.42	2.65	-1.5	-	0.24
	40	1.00	+0.66	-0.03	0.00	+0.63	1.70	3.33	-0.9	-	0.32
	50	1.25	+0.79	-0.03	0.00	+0.76	1.89	3.90	-0.3	-	0.39
	60	1.50	+0.90	-0.03	0.00	+0.87	2.02	4.39	0.2	-	0.46

* This is the channel contribution, with the transient correction:
(i) = steady-state contribution from the channel in its former position
(ii) = the recovery in temperature due to the presence of the river bar (t_1 assumed = 100 years)
(iii) = the contribution from the channel in its present position (for t_1 = 100 years).

** This represents the steady-state contribution from the channel in its present position and was the value used in the steady-state model. Comparison of values under (2) and (4) indicate the error involved in the steady-state model.

$$
\begin{aligned}
\theta(x,z,t) & \qquad\qquad\qquad\qquad\qquad\qquad\qquad\qquad (10)\\
= (\bar{T}_d-\bar{T}_s) & \left\{\left\{\psi\left[\frac{x-x_1}{z},m(t_1)\right] - \psi\left[\frac{x-(x_1+s)}{z},m(t_1)\right]\right\}\right.\\
& + \left\{\psi\left[\frac{x-x_2}{z},m(t_2-t_1)\right] - \psi\left[\frac{x-x_1}{z},m(t_2-t_1)\right]\right\}\\
& \left. + \left\{\psi\left[\frac{x-x_3}{z},m(t-t_2)\right] - \psi\left[\frac{x-x_2}{z},m(t-t_2)\right]\right\}\right\}\\
- (\bar{T}_d-\bar{T}_s) & \left\{\left\{\psi\left[\frac{x-x_1}{z},m(t-t_1)\right] - \psi\left[\frac{x-(x_1+s)}{z},m(t-t_1)\right]\right\}\right.\\
& \left. + \left\{\psi\left[\frac{x-x_2}{z},m(t-t_2)\right] - \psi\left[\frac{x-x_1}{z},m(t-t_2)\right]\right\}\right\} \quad \text{for } t>t_2
\end{aligned}
$$

Here the notation $m(t - t_1)$ means the value of $m = z^2/4\alpha t$ for $t = (t - t_1)$, etc. The movement from $x = \infty$ to $x = x_1$ at time t_o (=0) is counted as stage 1. In equation (10) the first group of three terms represents the warming influence of the river (i.e., permafrost degradation), whereas the second group of two terms accounts for the thermal recovery behind the migrating river (i.e., permafrost aggradation).

Now, if n stages of movement are involved, equation (10) becomes:

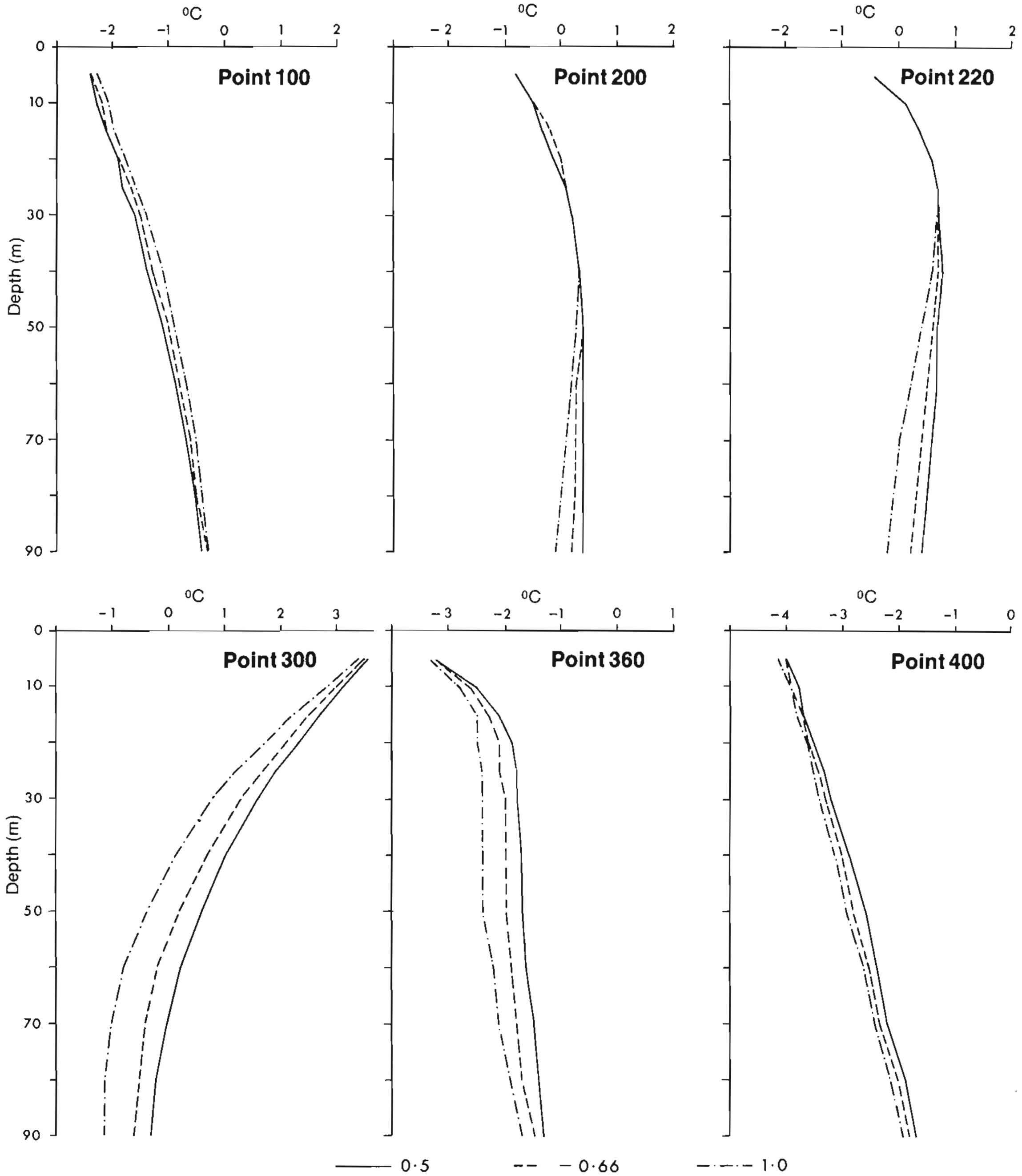

Figure 18. Effect of river shifting rate on magnitude of thermal disturbance.

$$\begin{aligned}
&\theta(x,z,t) \\
&= (\bar{T}_d-\bar{T}_s)\Bigg\{\Big\{\psi\Big[\frac{x-x_1}{z},m(t_1)\Big] - \psi\Big[\frac{x-(x_1+s)}{z},m(t_1)\Big]\Big\} \\
&+ \sum_{i=2}^{n}\Big\{\psi\Big[\frac{x-x_i}{z},m(t_i-t_{i-1})\Big] - \psi\Big[\frac{x-x_{i-1}}{z},m(t_i-t_{i-1})\Big]\Big\}\Bigg\} \\
&- (\bar{T}_d-\bar{T}_s)\Bigg\{\Big\{\psi\Big[\frac{x-x_1}{z},m(t-t_1)\Big] - \psi\Big[\frac{x-(x_1+s)}{z},m(t-t_1)\Big]\Big\} \\
&+ \sum_{i=2}^{n-1}\Big\{\psi\Big[\frac{x-x_i}{z},m(t-t_i)\Big] - \psi\Big[\frac{x-x_{i-1}}{z},m(t-t_i)\Big]\Big\}\Bigg\} \\
&\qquad\qquad \text{for } t>t_{n-1} \qquad (11)
\end{aligned}$$

Equation (10) assumes that $\bar{T}_s$ is the same before and after the river has passed by; as already mentioned, this is not the case. However, this can be handled in the same way as a climatic change--viz. if $(\bar{T}_d - \bar{T}_{s_1}) = A_1$ for $t < t_1$, and some new $(\bar{T}_d - \bar{T}_{s_2}) = A_2$ for $t > t_1$, the result for a stationary strip is (Lachenbruch, 1957b, p. 1521):

$$\begin{aligned}
\theta(x,z,t) &= A_1\Big\{\psi\Big[\frac{x}{z},m(t)\Big] - \psi\Big[\frac{x-s}{z},m(t)\Big]\Big\} \qquad (12) \\
&+ (A_2-A_1)\Big\{\psi\Big[\frac{x}{z},m(t-t_1)\Big] - \psi\Big[\frac{x-s}{z},m(t-t_1)\Big]\Big\} \\
&\qquad \text{for } t>t_1
\end{aligned}$$

Further, if details of some climatic change also were known, these could be easily incorporated.

Finally, the effect of a number of stages of movement, combined with that of a number of surface temperature changes, can be obtained by combining equations (11) and (12). The second group of terms in equation (11) is expanded to incorporate the effects of a varying $\bar{T}_s$. A computer program was written to handle this solution.

The case with v = 0.5 m per year will serve to illustrate some of the computational details. The total extent of the wave is 350 m; 230 m is the distance the river has shifted in 460 years. A total of 35 ten-metre strips represents the wave (Fig. 15). The strip between 230 m and 220 m corresponds to a surface temperature drop of 4.0°C; from 220 m back to 190 m, each strip corresponds to (1.2/3) = 0.4°C; and from 190 m back to 0 m, each 10 m spans a temperature drop of (3.0/19) = 0.16°C. The problem consists of evaluating the temperature effect of each of these ten-metre strips at all (x, z)'s in the cross-sectional space, using the appropriate values of $\bar{T}_s$ and t. As an example, the thermal history of the strip between 140 m and 150 m is as follows (remembering that 10 m corresponds to 20 years, for v = 0.5 m per year): for the first 60 years, $\bar{T}_s = -4.2°C$ as the river approaches; for the next 240 years, $\bar{T}_s = 4.0°C$ as the river wave "passes by"; thereafter the surface temperature declines, in 20-year intervals, to the present -1.8°C. In this way the changing thermal regime related to the vegetation succession is accounted for. The present temperature field reflects the sum total of all these stages. The procedure is repeated for all the strips.

Before the predicted temperatures can be compared to observed values, one final effect has to be included--that of the snowbank zone on the slip-off slope. This can be incorporated as a minor wave following the river disturbance, with the same migration speed, v (Fig. 15). The shape of this wave was formed in accordance with temperature measurements; its amplitude is about 1.0°C, and surface temperatures in this zone are close to 0°C.

Figure 16 presents the computed present-day permafrost configuration in the vicinity of the migrating river, using v = 0.5 m/year and $\alpha = 21.9\ m^2/year = 0.06\ m^2/day$. Observed and predicted temperatures for the borehole locations are given in Table 12. The initial condition, prior to migration, is partially shown (Fig. 16). The final computed temperatures also reflect the addition of the thermal contribution of all other water bodies in the area. The overall pattern computed by the model generally is consistent with the observed configuration; it is also similar to that described in other field studies (Péwé, 1965; J.R. Williams, 1970). Some of the details of the solution, however, are not perfect.

The permafrost history can be described in general terms as follows. In the initial position a curtain-shaped through-talik exists beneath the river, with the permafrost boundaries plunging steeply downwards. As the river migrates across the land surface this causes the talik to migrate also. In Figure 16, the isotherms beneath the river and in the lee of its migration slant away from the surface. This is due to the lag, with depth, in the penetration of the temperature disturbance as it moves across the surface. The lag in talik formation is shown by the bulge in the isotherms beneath the cutbank (cf. the permafrost boundaries in Fig. 16). There is a remarkable uniformity of temperature at depth beneath the river itself. Following the river disturbance, surface temperatures on the slip-off slope gradually decrease and permafrost wedges back in again. This configuration is a function of the decreasing surface temperatures away from the river and the lag in temperature recovery following disturbance. Temperatures on the cutbank side of the river are colder than those under the slip-off slope.

On the cutbank side, agreement between observed and predicted values is very close; the maximum deviation is 0.4°C and agreement is typically within 0.1° to 0.3°C. The predicted values are generally slightly colder except close to the river itself. Deviations ($T_{obs} - T_{pre}$) increase negatively with depth, and the predicted permafrost thickness of 80 to 90 m compares to the 60 to 70 m value of the steady-state model.

On the slip-off slope side the thermal regime is again quite faithfully reproduced in the zone within 100 m of the river. Here, agreement is typically within 0.2°C to 0.5°C, with the predicted values generally colder and permafrost thicknesses over-estimated. At site 6-5, for example, the predicted thickness of 13 m compares to the measured value of 10 m. With the snowbank effect included, a zone of permafrost degradation is produced under the slip-off slope (shown in detail in Fig. 17) which is very similar to that observed in the field (Fig. 7). In this vicinity the overall pattern of isotherms is disrupted with deeper, warmer isotherms being drawn towards the surface. This is the characteristic configuration around a talik (cf. the isotherms beneath the river in Fig. 16). As the distance from the river increases, however, the agreement progressively deteriorates with deviations becoming greater than -1°C. At site 6-12, for example, about 140 m from the river, the deviations range up to -1.4°C, and the predicted permafrost thickness of 50 m compares to the measured value of 22.5 m. It is in this zone that the model is least satisfactory.

What could account for the errors apparent in the model? Agreement is good in surface layers, with deviations becoming larger at depth. It is possible that the surface boundary conditions are in error, but the factor of climatic change is unknown. It is possible that the value for Gg might be too small, although this error is unlikely to be large and would not significantly offset the trend of larger negative deviations with depth. It is possible that some aspect(s) of the migration history has not been adequately reproduced in the model. This is a complex problem and cannot be readily evaluated. If the river did pause at some stage, however, this presumably would cause warmer temperatures in that vicinity. The probable explanation of the errors is the exclusion of the latent heat term in the present solution. If it were included, temperatures beneath the slip-off slope would not adjust so quickly to the surface cooling and therefore would be warmer.

Finally, mention should be made of the possibility of heat transfer by groundwater flow through taliks which could conceivably be quite considerable (e. g., Leschikov and Zarubin, 1967).

The thermal effect of the river on the surrounding ground temperatures depends not only on the strength of the source but also on the length of time available to the thermal processes. The latter is a function of the rate of migration, v. The effect of varying the speed of migration is illustrated in Figure 18. As v is increased to 1 m per year, temperatures on the cutbank side (point 400) tend to be slightly cooler, while those under the slip-off slope (point 100) are a little warmer. For slower migration rates the river will have been near to the present cutbank for a longer period, and thus the warming effect consequently will be greater. Faster migration rates for the slip-off slope will mean there has been less time for temperature recovery (i. e., cooling). The ground around the river shows the greatest effects of a changing rate of migration (points 220, 300, 360). If the river migrates more rapidly there is less time for the warming wave to penetrate to depth. Because of the increasing lag in penetration with depth, the effects of varying v are more pronounced at greater depths. In a model using a constant rate of migration, a value of about 0. 5 m per year gives the best overall results. There seems to be no way to establish the plausibility of this overall value, using vegetation data, at least. Additional field information is necessary to date certain features which might provide more information about the actual migration history.

Rate of Channel Shifting Inferred from Temperature Borehole Data

Another interesting problem that can be analyzed through heat conduction theory is the inverse of the one above--viz. using present-day data from temperature boreholes near a shifting channel, is it possible to determine something of the migration history? This can be approached using a simple single-step version of the transient model.

The back of the slip-off slope is marked by a fossil cutbank (Fig. 3). Assuming that this delineates some "initial" position of the channel, the steady-state temperature effect can be computed for the channel occupying this position using equation (4), with m = 0 ($t = \infty$). For site 2-6 ($\bar{T}_s = -3.0^oC$, $\bar{T}_d - \bar{T}_s = 4.0 - (-3.0) = 7.0^oC$) this works out to be 1.26^oC at a depth, z, of 15 m (x = 22 m, the distance from the fossil cutbank). With this contribution the predicted temperature is then -1.1^oC, which is 0.5^oC warmer than that now observed (Table 13). In other words, since the river began to shift, the ground at this location has undergone about 0.5^oC of cooling. As a first approximation the channel can be assumed to have shifted to its present position (a distance of 75 m) in a single step some t years ago and has been replaced by a bar 75 m wide. A weighted overall mean of -0.5^oC is used for the surface temperature of the bar.

The problem, then, is to calculate how long ago a step-change would have had to occur to produce the observed recovery of temperature at site 2-6. Equation (4) was solved for t, with the bar replacing the river as the strip-shaped disturbance. Thus ($\bar{T}_d - \bar{T}_s$) is -4.5^oC (i. e., the bar is 4.5^oC colder than the river which it has replaced), and the thermal disturbance, θ, is -0.5^oC (i. e., the 0.5^oC of cooling since the river began to shift). The width of the bar, s, is 75 m, and the borehole co-ordinates (x, z) are 22 m and 15 m respectively. If a state of equilibrium is disturbed, the rate at which a new equilibrium is established is controlled by the thermal diffusivity of the ground. With $\alpha = 18.3$ m^2 per year (a value somewhat lower than that for frozen ground, since the surface of the bar was initially unfrozen when beneath the river), t turns out to be 110 years; with $\alpha = 21.9$ m^2 per year, t is 93 years.

Measurements on aerial photographs taken in 1935, 1950, and 1967 indicate that the cutbank opposite retreated about 35 m during these 32 years, or about 1. 1 m per year. Between 1935 and 1950 the retreat was about 17 m, and between 1950 and 1967 it was about 18 m. Measurements over three summers (1969 to 1971) from some 40 stakes along the cutbank section opposite give an average retreat of about 2. 2 m (standard deviation = ± 0.9 m)[1]. Consequently, a predicted retreat of 75 m in about 100 years certainly seems to be realistic.

In treating the thermal history in such a simple fashion the thermal effect due to the river occupying its present position for the past 100 years has been ignored. For site 2-6, which is 97 m from the present channel, this thermal contribution is, in fact, very small for all z (Table 13).

To check the consistency of this model, the transient correction was applied to the prediction of temperatures at other boreholes along transect 2 (Fig. 1). Using a value of 21. 9 m^2 per year for thermal diffusivity, α, and 93 years for the migration period, t, θ was calculated from equation (4) for specified (x, z)'s. Agreement between observed and predicted temperatures was excellent (Table 13). It appears that this technique can provide a reasonable estimate of the rate of lateral migration from borehole data near the channel. The boreholes should be located off the river bar.

CONCLUSIONS

Although climate is basic to the formation of permafrost, this study has demonstrated that local factors are responsible for wide variations in permafrost conditions over a small area.

The combined thermal influence of the large number of surface water bodies is manifested in thinner permafrost than observed on the nearby tundra. Further, the permafrost of the Mackenzie Delta must be of

[1]Since the field observation period does not span three full ice-free seasons, but only about two-and-one-half, the current average rate of retreat possibly is closer to 2. 2/2. 5, or about 0. 9 m per year.

of a highly discontinuous nature, since permafrost is absent beneath the larger lakes and channels. It is estimated that in the absence of water bodies permafrost would be about two or three times thicker in the Delta.

In addition to spatial variations, geomorphological and biological evidence shows that surface conditions also are changing with time. The study area contains a landform assemblage that is typical of the Delta environment. A major distributary undergoing lateral migration is actively cutting into a frozen, mature, spruce-covered surface on the outside bends of meanders, with consequent degradation of permafrost. New alluvium is deposited on slip-off slopes, and permafrost forms there. The progressive reduction in ground temperatures in the lee of river migration is partly related to the process of thermal recovery following the river disturbance, and partly is a result of the insulating effect of the developing vegetation. Superimposed on these two trends is the permafrost degradation associated with the snowbank. The consistency of this interpretation of ground temperature variations was demonstrated through the framework of simple heat conduction theory.

In spite of a few large deviations between observed and predicted temperatures, the results obtained from the heat conduction models are interpreted as confirmation of their general validity. Mean annual surface temperatures, including those for water bodies, are important input parameters to these models. Once their values are established they can be used, together with a knowledge of the geothermal gradient, to predict the ground temperature field from the equations presented for areas where no measurements are available.

Specifically, the major conclusions of the study are as follows:

1. Temperature borehole data indicate that permafrost thicknesses in the order of 60 to 80 m are broadly representative of mature, spruce-covered areas. The greatest thicknesses are at sites more distant from water bodies; calculations show that the maximum thickness in the area is about 100 m, a value in good agreement with other measurements from the Mackenzie Delta.
2. Permafrost is much thinner (less than 10 m) in areas of recently deposited alluvium. Values were determined from temperature boreholes and resistivity sounding. On the slip-off slope, temperatures decrease and permafrost thickens with distance from the river. Permafrost may be completely absent in places on the slip-off slope with snow cover being the controlling factor.
3. Observed temperature gradients are about two or three times greater than provided for by the earth's geothermal gradient alone. The combined effect of the water bodies in the area was shown to account for the difference. In the absence of this effect, permafrost would be about 170 m thick at spruce-covered sites.
4. The mean annual air temperature in this area is -9° to -10° C. Estimated mean annual ground surface temperatures range from $+4.0^{\circ}C$ (river bottom), $+3.2^{\circ}C$ (lakes), -1.0° to $-1.5^{\circ}C$ (bare ground on slip-off slope), $-3.0^{\circ}C$ (willow-alder association), to $-4.2^{\circ}C$ (spruce forest). Borehole temperatures at 15 m range from about $+0.4^{\circ}C$ to $-3.5^{\circ}C$.
5. Through the framework of heat conduction theory a consistent explanation of permafrost in terms of local physical factors was developed and served to confirm the general validity of the hypotheses advanced. The calculated variations in permafrost distribution compare well with field measurements and, in a qualitative sense, with those reported in the literature.
6. The heat conduction models can be ultimately confirmed only by obtaining ground temperature data from depths to 60 to 80 m.
7. Application of the steady-state model to sites in stable areas yielded predicted results which are in excellent agreement with field observations. This indicates that with a knowledge of mean annual ground surface and water temperatures and the earth's geothermal gradient, the ground temperature field can be reliably calculated. A computer program was written for this purpose.
8. For areas of geomorphic change, the steady-state model is not satisfactory. The use of a temperature wave to simulate changing thermal conditions associated with channel migration permitted a realistic simulation. This model yielded generally satisfactory prediction of the thermal disturbance due to channel shifting. Beneath the bar the predicted regime duplicated the observed pattern well; inclusion of the thermal wave associated with snowbank migration produced a zone of permafrost degradation as observed in the field. The deterioration in predicted values at distances away from the channel possibly results from the failure to include the latent heat term in the equations. This cannot be incorporated in any simple way.
9. A smiple, single-step transient analysis of temperature borehole data was used to estimate the rate of channel shifting and yielded a predicted value in close agreement with that determined from actual field measurements.
10. Calculations carried out indicate that through-taliks exist beneath channels and lakes that are wider than 80 to 100 m.
11. On the slip-off slope variations in snow accumulation produce significant ground temperature variations. It is concluded that snow cover is a permafrost-controlling factor in this locality. Beneath the snowbank zone a talik has formed as a result of the insulating effects of deep snow.

ACKNOWLEDGMENTS

Financial support for this study was received from the Geological Survey of Canada and from research grants given to J.R. Mackay, Department of Geography, University of British Columbia, by the Department of Indian and Northern Affairs and the National Research Council of Canada. I extend my thanks to these agencies and to Dr. Mackay, and also to the Inuvik Research Laboratory for the help they provided. I also extend my appreciation to Imperial Oil Limited for generously providing travel facilities on several occasions, and to Gulf Oil Limited for co-operation in drilling a number of boreholes.

I also wish to express by gratitude to the many individuals from whom I received help and encouragement during the research and the preparation of this paper, particularly P. A. Carr and R. A. O'Neil who carefully read and commented on the manuscript.

Lastly, I wish to thank Dean R. Wendt and Carleton University for a grant to cover the costs of preparing the manuscript, and April McIlhagga, Bonnie Gray, and Stefan Palko for their assistance in the preparation of the original thesis from which this paper has been prepared.

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APPENDIX

Glossary of Terms

depth of zero annual amplitude	The depth to which seasonal temperature fluctuations extend into the ground. Beneath this, temperature remains constant year-round.
frost table	The surface which represents the level, at any time in spring and summer, to which thawing of the seasonally frozen ground has penetrated.
geothermal gradient	The increase of temperature with depth in the earth due to the heat received from sources within the earth.
pereletok	A frozen layer at the base of the active layer which remains unthawed for one or two summers.
permafrost table	The surface which represents the upper limit of permafrost.
talik	An unfrozen portion within a body of permafrost. It commonly implies thawed ground that was probably permafrost at some time.
-through-talik	A thawed zone that perforates the permafrost.
-pseudo-talik	A thawed zone that does not perforate the permafrost. The permafrost table is locally depressed, while an upward indentation is formed in the base of permafrost.
terrain segment	A portion of the surface area which can be characterized by a degree of homogeneity in its physical composition (nature of surface cover, topographic position).