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NAME OF AUTHOR/NOM DE L'AUTEUR MARGARET BURGESS

TITLE OF THESIS/TITRE DE LA THÈSE ANALYSIS OF THE GROUND THERMAL REGIME
AT NORMAN WELLS AND FORT GOOD HOPE;
1971-1974

UNIVERSITY/UNIVERSITÉ CARLETON UNIVERSITY

DEGREE FOR WHICH THESIS WAS PRESENTED/
GRADE POUR LEQUEL CETTE THÈSE FUT PRÉSENTÉE MASTER OF ARTS IN GEOGRAPHY

YEAR THIS DEGREE CONFERRED/ANNÉE D'OBTENTION DE CE GRADE 1983

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ANALYSIS OF THE GROUND THERMAL

REGIME AT NORMAN WELLS AND

FORT GOOD HOPE: 1971-1974

by

Margaret Burgess, B. Sc.

A thesis submitted to the Faculty of Graduate
Studies and Research in partial fulfillment
of the requirements for the degree of Master

of Arts

in Geography

Department of Geography

Carleton University

Ottawa, Ontario

August 1983

The undersigned recommend to the Faculty of Graduate
Studies and Research acceptance of the thesis

"ANALYSIS OF THE GROUND THERMAL REGIME AT NORMAN WELLS AND
FORT GOOD HOPE: 1971-1974"

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September 29, 1983

ABSTRACT

Shallow ground temperature data from 11 sites in the Norman Wells - Fort Good Hope region are analyzed to characterize the ground thermal regime over the 1971-1974 observation period. Multiple harmonic Fourier series interpolations are performed to regularly generate temperatures from the discontinuous record, for characterisation of annual ground temperature variations.. Numerical techniques to calculate thermal diffusivity from the temperature data are explored. These are found unsuitable, and thermal properties are calculated from physical properties. Numerical simulations of microclimatic and ground thermal regimes indicate that the differences in temperatures from site to site can be largely attributed to lithology differences and to variations in surface moisture and snowcover conditions. Ground temperature warming observed at most sites probably arises from surface disturbances related to the borehole drilling program. In future monitoring programs of shallow ground thermal regimes, detailed measurements of soil physical properties and surface energy exchange components are recommended.

ACKNOWLEDGEMENTS

I am grateful to the Earth Physics Branch of Energy, Mines and Resources for granting my education leave and for financially assisting my work towards a Master's degree. Their support and understanding, when I realized I could not pursue my original thesis plan, were much needed and appreciated. I thank Alan Judge particularly for his support during those discouraging moments, and for suggesting the present analysis as an alternate thesis topic. I thank my thesis supervisor, Mike Smith, for his guiding light; his concern, suggestions and constant feedback at our regular meetings were invaluable.

Dan Riseborough provided help during the numerical simulation phase. Carole Brown, Chris Lacasse and Micheline Whissell assisted me in my struggle, at times very frustrating, to produce a manuscript on the word processor. Suzanne Cumin drafted many figures; Vic Allen provided instructions and assistance in operating the HP 85 graphics unit for drawing the remaining figures. Ken Torrance and Al Taylor kindly read the manuscript and made several useful comments. Many thanks to all these people.

All my thanks and love to Stephen Barrette for his constant moral, emotional and domestic support during the last two years. Life with someone never quite able to get a thesis off their mind can be exacting.

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CHAPTER 1

INTRODUCTION

In the early 1970's, an intensive studies program was undertaken in the Mackenzie Valley and Northern Yukon to examine the possible environmental and social effects of a proposed northern gas pipeline. The research was carried out under the Environmental-Social Program (ESP) of the Northern Pipelines Task Force on Northern Oil Development, Government of Canada. The Earth Physics Branch of the Department of Energy, Mines and Resources participated in the environmental studies program through its investigation of geothermal conditions in the ground surface layer. This research included the determination of geothermal heat flows, ground surface temperatures, permafrost thickness and thermal properties of frozen materials (see Judge, 1973a; 1975). In addition, the Geological Survey of Canada conducted studies under the ESP into terrain sensitivity and engineering and physical properties of permafrost materials (see Kurfurst, 1973; Isaacs, 1974).

These investigations necessitated the drilling of shallow boreholes (up to 60 m in depth) for observations and measurements of ground temperatures and soil physical and thermal properties. Ground temperatures were monitored for up to 2.5 years at eleven sites in the Norman Wells and Fort Good Hope region (Figure 1.1). The sites represent a variety of ground surface and lithological conditions. Only the first year's data were analysed in the above reports. The 2.5 year temperature record presents a unique opportunity to make not only general observations of the thermal regime (e.g. depth of the active layer, mean annual surface temperature, presence or absence of permafrost), but also to analyse and compare ground thermal regimes at the various sites.

Figure 1.1 Site location map

The research undertaken in this thesis attempts to resolve the effects of subsurface lithology, site specific surface conditions (e.g. vegetation, wetness, snow cover), natural climatic trends over the observation period, and man-induced or other disturbances, on the observed ground thermal regime at each site. In addition, an investigation of the interaction of these factors should be useful in understanding the effects of climatic change on ground temperatures. Large scale changes in air temperatures are not translated to the ground thermal regime in a straightforward way. Air-ground interactions are complex, and modulating effects vary according to microclimate and lithology. A knowledge of the behaviour of ground thermal regimes in response to natural or man-induced surface changes has important implications, however, for northern development (Federal Environmental Assessment Review Process, 1981; Burn et al., 1982).

The following quote from Judge (1975, p.2) highlights the three main thrusts of the thesis research:

" Ground temperature information, from both disturbed and undisturbed sites, aids in assessing the maximum effect of surface changes on ground temperatures, and measurements of thermal properties over a range of temperatures aids in establishing the rate at which surface changes are translated to the subsurface. Numerical or analytical simulations aid in establishing both these rates and the natural rates of change due to changes in climate."

Chapter 2 provides the background information to the study, outlining the general nature of ground thermal regimes and the regional setting of the study area. The first component of the research consists of synthesizing the ground temperature information to characterize the ground thermal regime at each site (Chapter 3). The ground temperature data base consists of temperature logs recorded for up to 2.5 years at eleven sites; however, the data base required much work before a detailed analysis could begin. An

overall quality control was necessary to ensure that erroneous values were removed and that the data set was complete. Temperature logs were not recorded at regular intervals, and in several cases gaps of up to 5 months exist in the record. An analysis of several curve fitting routines was undertaken, so that gaps could be filled and temperatures interpolated at regular intervals. Programs to calculate and plot running mean temperatures, and to contour geotherms (ground temperature isotherms over time), were developed to compare sites over the observation period.

The second component of the research involves the categorization of the soil physical and thermal properties for each site (Chapter 4). Geotechnical measurements available from the boreholes served as a basis for determining physical properties of soil layers. Analyses of the thermal properties of soil layers focused on the thermal diffusivity, which is most useful in establishing the manner in which surface temperature changes are transferred to the subsurface. Several methods, all of which rely on the use of the temperature record over time and with depth, were explored for calculation of diffusivities. Thermal conductivities and heat capacities were calculated from the soil physical properties for a range of temperatures.

The soil physical and thermal properties were used, along with atmospheric data from the Atmospheric Environment Service's weather station at Norman Wells, in numerical simulations of microclimatic and ground thermal regimes in the third component of the research (Chapter 5). A sensitivity analysis, using the computer simulation of Smith (1977), was performed to examine the effect of lithology, site specific characteristics and climatic variation on the ground thermal regime. The analysis was divided into three parts:

- i) effect of lithology: using climatic normals from the region, and imposing the same surface conditions at each site, the thermal response of different soil lithologies was simulated by running a uniform climate until equilibrium profiles were established.
- ii) effect of natural climatic variations: starting with the above stabilized profiles, the climatic conditions over the observation period were imposed.
- iii) effect of microclimate: the ground thermal response to variations in surface conditions (in particular site wetness and snowcover) was determined for climatic normal conditions.

Several difficulties with the numerical simulation were encountered and overcome before the modelling proceeded. These problems served to stress how a lack of sufficiently detailed descriptions of site conditions limited the simulation of representative conditions. The analyses also suggested additional studies for future research.

RELATION TO NORTHERN DEVELOPMENT

Judge (1975) recommended that geothermal observation programs in the Mackenzie Valley be repeated at 5-year intervals to determine long term trends resulting from small changes in surface temperature. The Berger Inquiry (Berger, 1977) examined the environmental, social, economic, and geotechnical considerations of the Mackenzie Valley project. The preservation of the integrity of the thermal regime and minimisation of disturbances to surrounding terrain were of prime concern (Williams, 1979). Berger recommended

a 10 year moratorium on development. His report was not endorsed by the Government of Canada.

In the late 1970's, Esso and Interprovincial Pipe Line Ltd proposed the Norman Wells Oilfield Expansion and Pipeline Project. Production of the Norman Wells reservoir would be increased to 4500 m³/day by waterflooding techniques. The project would require 175 new oil and water injection wells, 6 artificial islands in the Mackenzie River, oil gathering facilities, a central processing plant (natural gas would be returned to wells to assist in lifting the oil) and an 866 km pipeline from Norman Wells to Zama, Alberta. The pipeline, 324 mm in diameter with a 5000 m³/day capacity, would be buried throughout its length at a minimum depth of 1 m and would carry products at 'near ambient' ground temperatures.

The project, and the pipeline in particular, presented several possibilities for disturbances to the ground thermal regime. The following concerns were stressed by the Federal Environment Assessment Panel (FEARO, 1981):

- 1) by operating at ground temperatures, cooling in cold areas could lead to freezing and heaving in unfrozen ground downstream
- 2) or vice versa, heating could lead to thawing and settlement
- 3) the design for maximum thaw settlement did not take into account the compounding effects by surface changes, climatic changes and forest fires
- 4) the possibility of the pipe jacking up with repeated freezing and thawing.
- 5) the ground thermal regime analysis was cursory and insufficient to determine the likelihood or possible magnitude of problems

Construction of the Norman Wells Oilfield Expansion Project began in early 1983 (Information Pipeline, Jan. 1983). The Department of Indian Affairs and Northern Development is planning a research and monitoring program for the project - a component of which would involve the evaluation of the

impact of the pipeline on permafrost (A.S. Judge, pers. comm.). The geothermal studies program that ended abruptly in the mid 70's might thus be extended, and investigations of the interaction of climate, microclimate, surface disturbances and ground thermal regimes once more pursued.

The present thesis research was designed to analyse recorded ground thermal responses for several ground surface and subsurface conditions, over a 2.5-year period, and hopefully contribute to understanding the interaction of climate, microclimate and ground thermal regime. In addition, the research results can also be applied to the design of future research and monitoring programs. In the past these programs have either not lasted long enough, or not collected enough information (both in terms of number of parameters measured and frequency of measurements).¹ Monitoring programs, involving shallow ground thermal regimes, need a much greater emphasis than has been allotted in the past, on gathering of information related to all the components of the surface energy exchange throughout the year.

1. The resulting incomplete data sets are due more often than not to financial constraints, rather than to a lack of knowledge of the important factors to be measured.

CHAPTER 2

BACKGROUND TO THE PRESENT STUDY

2.1 FACTORS INFLUENCING THE GROUND THERMAL REGIME

Unlike the term permafrost, there is no concise definition of the phrase 'ground thermal regime'. Most books on cold regions engineering and geotechnical studies devote chapters or sections to describe the nature of ground thermal regimes in permafrost areas (e.g. Judge, 1973a; Andersland and Anderson, 1978; Johnston, 1981; Lunardini, 1981). The words 'ground thermal regime' bring to mind notions of heat transfer within, and temperature conditions and thermal properties of, subsurface materials. In earth materials, heat may be transferred by radiation, conduction and mass transfer processes. Radiative heating and cooling are only important at the earth's surface. Mass transfer processes (convection) can play a role where ground water is present and moisture movement is possible. The principal method for heat transfer within the ground is conduction.

Temperatures within the ground are bounded on the one hand by the surface temperature and, on the other, by the geothermal heat flux. The geothermal heat flow - resulting largely from heat released by radioactive decay within the earth, and perhaps other sources such as residual heat from the formation of the earth - produces a geothermal gradient. Temperatures increase with depth at an average world value of approximately $25-30 \text{ mKm}^{-1}$. If the earth's surface temperature were constant, the rate at which temperature increased with depth would only be a function of the geothermal heat flux of the area and the thermal conductivity of the earth materials.

These are the conditions for steady state heat transfer. However, the surface temperature, the upper boundary condition, fluctuates both daily and seasonally. These temperature changes are felt at depth, with an attenuation in the amplitude and a lag in the phase. Under such conditions of transient heat transfer, the depth and nature of penetration of the surface temperature wave depend on the thermal diffusivity of the earth materials - the property that governs the rate at which a material responds to changes at the surface.

On an annual basis, a mean temperature can be determined for the ground surface. The depth to which the annual surface temperature wave penetrates is known as the level of zero annual amplitude. This level can be as much as 15 to 30 m, depending on the nature of the material. Within the zone of annual oscillations, the minimum and maximum temperatures reached at each level define the ground temperature envelope. These characteristics of the ground thermal regime are illustrated schematically in Figure 2.1, for the case of a permafrost environment. Permafrost conditions occur where earth material has been below 0°C for more than two winters plus the intervening summer. The layer above permafrost that freezes each winter and thaws each summer is the active layer.

The thermal regime in permafrost soils is often complicated by the presence of water. Heat transfer processes then involve the absorption or release of latent heat, because of the phase change of water. These phase transformations retard the rate at which ground temperatures change. In addition, the thermal properties of frozen wet soils are substantially different from those of unfrozen wet soils (see, for example, Farouki, 1981). The thermal conductivity of ice is roughly four times that of water, while its heat capacity is less than half that of water. Further, soil water changes phase over a temperature range rather than solely at 0°C , and each soil has its own freezing characteristic curve. In fine-grained soils, large volumes

Figure 2.1 Diagrammatic sketch of the thermal regime in permafrost

of water can remain unfrozen at very low temperatures. The possibility of moisture flow and convective heat transfer thus arises both in the unfrozen and 'frozen' ground.

Lunardini (1981) divides the permafrost profile into several energy balance layers (Figure 2.2). Within the permafrost layer there are several subdivisions, depending largely on temperature and, hence, on the state of soil moisture. Ground temperatures at depth below the permafrost layer are functions of the lithosphere, where the geothermal heat flux varies according to the age and tectonic evolution of the region. In studies of the shallow ground thermal regime - depths less than 100 m and generally extending to the level of zero annual amplitude - the energy balance at the surface of the earth is of prime importance. This surface energy balance is a function of the interaction of the climate and the local microclimatic setting, and determines the ground surface temperature. The sun's radiation is the fundamental factor governing the energy balance, and the geothermal heat flux accounts for only a small percentage of the energy exchange.

Several complex energy transfer processes occur between the ground and the atmosphere. These result from the following physical processes (Lunardini, 1981):

- 1) Surface radiation: the net sum of the outgoing and incoming short and long wave radiation
- 2) Atmospheric convection: turbulent heat transfer between the surface layer and atmosphere
- 3) Moisture flow: evaporation and condensation; movement of moisture in ground layers
- 4) Phase change in ground layers
- 5) Geothermal heat transfer
- 6) Conduction and convection in ground layers

<u>Layer</u>	<u>Description of layer</u>
1	Portion of lower atmosphere adjacent to surface layer
2	Contact layer between soil and atmosphere, may include vegetation, snow, etc. This layer is difficult to specify exactly for all possibilities.
3	Thin layer of soil adjacent to contact layer, arbitrarily 10 cm
4	When added to layer 3, this constitutes the active layer
5	Region of frozen soil in which water may be liquid
6	Frozen zone with no liquid water; 3, 4, 5 and 6 constitute the soil layer in which temperature oscillations occur
7	Frozen zone with no liquid water
8	Frozen zone with liquid water, same criteria as layer 5.
	Layers 5, 6, 7 and 8 constitute the thickness of permafrost
9	Thawed layer below permafrost - the lithosphere

FIGURE 2.2 POSSIBLE ENERGY BALANCE LAYERS IN PERMAFROST

(after LUNARDINI, 1981)

At any given moment the surface energy balance can be written as:

$$Q^* + Q_H + Q_E + Q_S = 0$$

where Q^* = net radiation flux

Q_H = sensible heat flux (convective component of balance)

Q_E = latent heat flux (evaporation, condensation)

Q_S = soil heat flux

Thorough discussions of these surface energy balance components are provided in Geiger(1975), Oke(1978) and Lunardini(1981). Net radiation is made up of incoming solar and emitted earth radiation. Incoming solar radiation is partially reflected upon contact with the surface, the proportion being determined by the surface albedo, which varies according to vegetation type, snow cover, surface wetness, etc. The amount of incident solar radiation varies with latitude, time of year, and angle and orientation of slope. Longwave radiation is emitted by the earth according to its absolute temperature and emissivity. Much of this longwave radiation is absorbed and re-radiated back to earth by clouds, water vapour and carbon dioxide.

The sensible and latent heat fluxes are both convective transports of energy via eddy motion. The aerodynamic properties of the surface determine, under given weather conditions, the form of the vertical wind profile and the state of turbulence of the air, which are central to the magnitude of Q_H and Q_E . Soil and air relative humidities and temperatures define the vapour concentration gradient. Given the existence of a gradient, and the availability of water and energy, evaporation can occur.

Most of the surface conditions influencing the surface energy exchange, e.g., vegetation type, wetness, snowcover, albedo and aerodynamic roughness, are subject to seasonal variation. Spatial variations in these conditions also occur over short distances. Different microclimatic regimes

give rise to differing ground thermal regimes and thus to lateral heat flows below the surface. The ground thermal regime at any location is thus not only a function of the energy balance at the site, but at neighbouring locations as well. For example, the presence of large bodies of water, such as rivers, lakes or ocean shorelines can have a considerable warming effect on ground temperatures in a permafrost environment (Lachenbruch et al., 1962; Smith, 1976).

On a longer term, the mean annual ground surface temperature may change in response to large scale variations in climate, such as occurred, for example, during the Pleistocene glaciations, the Climatic Optimum and the Little Ice Age. In areas of marginal or warm permafrost soils, slight changes in ground temperatures resulting from either natural climatic change, or from surface disturbances by man, can cause dramatic changes in soil thermal conditions. In ice-rich soils, a slight increase in mean temperature may result in an increase in the depth of summer thaw, the release of large volumes of water (more than would be normally contained in the unfrozen soil), and ultimately ground settlement. In unfrozen, frost-susceptible soils, a slight decrease in mean temperatures to below 0°C could create conditions conducive to ice lensing and frost heave (Williams, 1982).

In summary, studies of shallow ground thermal regimes in permafrost soils require a knowledge of the thermal properties of the earth materials over a range of positive and negative temperatures, of the annual ground temperature variations, of the geothermal heat flux, and of the many factors influencing the surface energy balance and its seasonal variation.

2.2 GEOHERMAL STUDIES IN THE MACKENZIE VALLEY

The summary of geothermal studies in the Mackenzie Valley presented below is based on the work by Judge (1973a; 1975). Mean annual surface temperatures are between 0°C and -3.0°C over much of the Valley and the ground thermal regime is characterized by the widespread presence of permafrost. Permafrost thicknesses vary widely, from 0 m in the south to over 500 m in the north, in response to the range in mean annual surface temperatures and ground temperature gradients. In the Fort Good Hope area, thicknesses range from 33 to 48 m and surface temperatures from 0.1 to -2.0°C ; in the Norman Wells area, thicknesses range from 26 to 70 m, and surface temperatures from -0.7 to -3.1°C . Both areas are within the discontinuous permafrost zone (Brown, 1967) and fall within Judge's thermal zone II (Figure 2.3) where mean annual surface temperatures are generally negative. Permafrost thickness in this zone is generally less than 50 m and does not exceed 80 m. The active layer thickness varies from 0.6 m beneath peat to 15 m in exposed bedrock.

Geothermal gradients in the Valley range from 10 mKm^{-1} (shield rocks) to 84 mKm^{-1} (shales). In the study area, average geothermal gradients are on the order of 60 mKm^{-1} . Thermal conductivities reported for frozen unconsolidated overburden materials average $1.9\text{ Wm}^{-1}\text{K}^{-1}$ for sands and $1.2\text{ Wm}^{-1}\text{K}^{-1}$ for clay. Judge found that frozen conductivities at temperatures below -2.5°C were generally 30% higher than unfrozen values. Bedrock conductivities range from $0.6\text{ Wm}^{-1}\text{K}^{-1}$ in poorly compacted shales to $3.5\text{ Wm}^{-1}\text{K}^{-1}$ in granites. Resultant terrestrial heat flow is fairly uniform along the Mackenzie Valley and high with respect to the rest of Canada. Published values for the Norman Wells region are on the order of 80 mWm^{-2} (Judge, 1973b).

Figure 2.3 Preliminary geothermal zones in the Mackenzie Valley (Judge, 1975)

Temperature profiles in the Mackenzie Valley suggest that temperature changes (as much as +2 K) have probably occurred as a result of climatic and vegetation changes in the last 100 years. These changes are evident as temperature inversions and as reductions in shallow ground temperature gradients. Similar changes have been observed in ground temperature profiles in southern Canada as a result of climate warming from the late 19th century to the 1940's (see, for example, Cermak, 1971). Recent warming is also evident in ground temperature profiles from northern Alaska (Lachenbruch et al., 1966; Lachenbruch et al., 1982).

Man's presence in the north may also induce changes in surface conditions, effectively changing the microclimatic and ground thermal regimes. The effects of man-induced disturbances become superimposed on natural climate trends, perhaps making their distinction difficult. Climate varies naturally from year to year, as the plot of mean annual air temperatures at Norman Wells for 1945 to 1980 shows (Figure 2.4). The general technique for removing short term variability is the use of running means. Five-year running means of air temperatures are also plotted in Figure 2.4, and indicate an overall decrease in mean temperature of ~1 K from the 1940's to 1960, a slight sinusoidal increase of less than 0.5 K from 1960 to the early 1970's and an increase of nearly 2 K in the 1970's.

Even when a knowledge of past climate change can be determined or a prediction of future climate change made, how these changes are transferred to, and modify, the ground surface and subsurface thermal regimes has been little investigated (Burn et al., 1982; Goodwin et al., in press; Smith and Riseborough, 1983). However an understanding of these interactions is important to future northern development, where structures such as pipelines may have a 30+ year lifespan, a period over which natural or man-induced climate changes could occur and profoundly change ground thermal conditions.

FIGURE 2.4 NORMAN WELLS AIR
TEMPERATURES SINCE 1945

2.3 GEOGRAPHIC SETTING

2.3.1 Climate

The Atmospheric Environment Branch of Environment Canada (AES) maintains weather stations at both Fort Good Hope and Norman Wells, recording temperature, precipitation, cloud cover, atmospheric pressure, and vapour pressure. In addition, the Norman Wells station records wind and radiation. Both stations have been in operation since the early 1940's. Fort Good Hope also ran another station from the turn of the century to 1966. The 30 year climate normals and monthly weather observations published by AES are used in the present research to provide data for microclimatic modelling (Chapter 5).

Climate plays a major role in determining the surface energy balance and thus the surface temperature of a site. Burns (1973; 1974) presents a detailed overview of the climate of the Mackenzie Valley and Beaufort Sea. The following summary of the climate of the Fort Good Hope and Norman Wells region is largely derived from his work. Burns has divided the Mackenzie Valley into climatic-zones with classifications based on physiography, vegetative cover and climatic characteristics. Both Norman Wells and Fort Good Hope fall within the Taiga Ecotone zone, whose climatic character is described below.

At 65°N latitude, the study sites border the land of the midnight sun and the polar night. In mid-winter the sun is only above the horizon for three hours a day, whereas in mid-summer there are nearly twenty-four hours of daylight. Despite the long hours of daylight, daily radiation received at the surface during the summer is only equivalent to that received in temperate latitudes because of the low angle of incidence of the sun's rays. In the winter months, radiation decreases to virtually zero in November through January. Mean annual solar radiation received on the ground is 56 MWhm^{-2} .

Cold continental Arctic air masses dominate in the winter, retreating in spring to give way to warm maritime Arctic air masses in summer. In winter, the Arctic inversion (air temperature increasing with height) arises from the negative radiation balance over the snow, due to the high albedo and low solar elevation. Stable inversions are persistent throughout the winter. In summer, radiation inversions are frequent at night but destroyed by heating during the daytime.

Annual precipitation is low, yet the ground surface can be very wet in summer, because drainage is impeded by the presence of the permafrost table. Further, although total annual precipitation is on the order of 35 cm W.E. (water equivalent), most of this falls in the summer. Norman Wells receives slightly more precipitation than Fort Good Hope. Mean annual air temperature is -8.0°C at Fort Good Hope and -6.3°C at Norman Wells. Their summer temperatures do not differ markedly; however, winters are colder at Fort Good Hope. Mean hourly wind speed decreases southward from the Arctic coast and averages annually 340 cm s^{-1} in the study area. Maximum means generally occur in the summer along the Mackenzie Valley, although the largest extremes occur in the winter. Near Norman Wells, wind distribution is strongly affected by topography, with a marked N-SE configuration corresponding to the valley orientation. Average cloud cover is at a minimum of 50% in winter and increases through spring and summer to reach a maximum of 75% in the fall (October-November). Clouds are predominantly low and middle types: stratus, stratocumulus, altostratus and altocumulus (Vowinkel, 1962).

2.3.2 Site Descriptions

Fort Good Hope and Norman Wells lie in the mid-Mackenzie Valley, in the Mackenzie Lowland Physiographic region. In the Norman Wells region the valley is rugged to the east side, due to the Norman Mountain Range, and broadens out to the northwest into a flat lake-strewn area. Norman Wells lies on the border of two forest regions: the predominantly forest zone to the south and the forest grassland zone to the north. Fort Good Hope lies within the latter zone. In the forest areas, alluvial flats are mainly covered by white spruce and balsam poplar; benches above the floodplains are covered by large areas of pines and aspen, with tamarack and black spruce in moist areas.

The seven Norman Wells (NW) sites cluster around the town, with distances between any two sites ranging from 1 to 25 km. The four Fort Good Hope (FGH) sites are located south of the town and spread over a wider area, with as much as 60 km between sites. The greatest distance between a FGH and NW site is ~175 km. The location of each site (latitude, longitude and elevation) and relevant drilling information are presented in Table 2.1. The Fort Good Hope sites were drilled in the summer of 1971 and logged until the summer of 1974; the Norman Wells sites were drilled in March 1972 and also logged until the summer of 1974. The temperature logs used in the present work do not include those recorded in the first few months after drilling, to ensure that any drilling disturbance had largely dissipated. This dissipation, for shallow holes where large amounts of latent heat are involved in refreezing, may take over a month (see, for example, the return to equilibrium of shallow subseabottom boreholes in the Beaufort Sea, in MacAulay et al., 1977). Thus, 2.5 years of data are available from the Fort Good Hope sites (except FGH-2 where the temperature cable failed after one year), and 2 years of data from the Norman Wells sites.

TABLE 2.1. SITE LOCATIONS AND DRILLING INFORMATION

SITE	LATITUDE (N)	LONGITUDE (W)	ELEVATION (m)	DEPTH (m)	DRILLING DATE	LOGGING PERIOD
FGH-1	65 44.2	128 41.4	30	30	30/ 6/71	12/71-6/74
FGH-2	65 47.7	129 25.4	91	10.7	27/ 7/71	12/71-1/73
FGH-3	65 46.9	128 51.9	30	12.2	4/ 8/71	12/71-6/74
FGH-4	66 7.3	129 36.3	69	23	24/ 8/71	12/71-6/74
NW-1	65 21.1	127 10.0	61	14	3/72	7/72-6/74
NW-2	65 15.1	127 0.2	76	14	3/72	7/72-6/74
NW-3	65 17.0	128 45.0	72	60	3/72	7/72-6/74
NW-4	65 20.1	127 4.1	61	14	3/72	7/72-6/74
NW-5	65 15.7	126 37.6	122	14	3/72	7/72-6/74
NW-6	65 13.5	127 7.0	168	14	3/72	7/72-6/74
NW-7	65 15.2	126 38.6	69	30	3/72	7/72-6/74

Logging intervals refer to period used for analysis in this report. Depth refers to interval monitored by temperature cable.

Brief descriptions of surface conditions at the drillsites were published in Judge (1973a) and Isaacs (1974). Table 2.2 summarizes the available information on vegetation, drainage, slope, and surface layer (blanks in the table indicate the information was not recorded). The sites were selected to represent a variety of surface and subsurface conditions. Most sites had to be cleared for the drilling operation (generally removal of trees and brush), and thus underwent some surface disturbance. FGH-1, FGH-2 and FGH-4 were located in an area disturbed by recent fire. Information from Forest Resources, Dept. of Indian Affairs and Northern Development in Fort Smith (R. Bailey, pers. comm.) indicates that two forest fires occurred in the vicinity of Fort Good Hope in 1969, and that the three sites were burned over by one of the two fires.

The comments in Table 2.2 reveal that nearly all sites, no doubt for reasons of access, particularly around Norman Wells, were located near some sort of terrain disturbance, e.g., a road, a right-of-way, a seismic line or a cut line. A survey of aerial photographs since 1952 indicates that these disturbances, for the most part, are not recent and have existed for the last 30 years - in other words some 20 years before drilling. NW-1, NW-2, NW-3, NW-4, NW-5, NW-7 and FGH-3 are located in either a medium or a dense forest. The remaining sites are located in areas with scattered tree cover.

Surficial mineral soils in the study area are glaciolacustrine and consist mostly of clayey silt with some silty clay and silty fine sand. In depressions, these fine grained sediments are overlain by peat, varying in depth from 0 to 3 m (Brown, 1965; Franci and Associates, 1974). Visible ice is common in the fine grained soils, with ice contents ranging from 30 to 60% by volume. The underlying bedrock is predominantly shales and sandstones. Simplified lithological logs for each study site, derived from the published data (Judge, 1973a; Isaacs, 1974) and used in this report, are shown in Figure 2.5.

TABLE 2.2 SITE DESCRIPTIONS

SITE	DRAINAGE	SLOPE	SURFACE LAYER	VEGETATION	COMMENTS
FGH-1	good	-	peat	hummocky: moderately dense trees destroyed by recent fire	15m above river: near eroding gully
FGH-2	poor	flat	peat	hummocky: tree cover destroyed by recent fire	just off a seismic line
FGH-3	good	flat	silt/organics	dense forest	cleared for drilling: enlarging previous clearing
FGH-4	poor	flat	sand	hummocky: thin forest destroyed by recent fire	100m from swampy lake
NW-1	good	slight	thin organic mat	dense black spruce forest	15m west of a seismic line
NW-2	-	-	silt/organics	dense black spruce	15m from old Canol road and a cut line
NW-3	fair	gentle	till/organics	black spruce	7m from old Mackenzie Valley pipeline right of way
NW-4	good	flat	thin peat	dense dwarf black spruce	5m from seismic line: 30m from lake: thermokarst area
NW-5	good	gentle	thin peat	dense dwarf black spruce	near winter road: 5m to north of an old seismic line
NW-6	poor	flat	thick peat	hummocky fenland with scattered trees: swamp	along old Canol road: standing water in summer
NW-7	good	gentle	sand & gravel with organics	dense black spruce	clearing at intersection of 2 seismic lines: 70m from large lake

(At all NW sites, trees and bush were bulldozed aside for drilling but the moss mat was left intact.)

FIGURE 2.5 a) LITHOLOGICAL UNITS

FORT GOOD HOPE SITES

FIGURE 2.5 b) LITHOLOGICAL UNITS
NORMAN WELLS SITES

CHAPTER 3

GROUND TEMPERATURES

3.1 INTRODUCTION

The ground temperature data base provides the main body of data for the thesis research. In this chapter the acquisition of the temperature data and the format of the data file are first briefly outlined. The preparation of the data set for analysis is discussed at some length: this treatment includes meticulous quality control and an investigation of several interpolating routines to select a suitable technique for calculating mean annual ground temperatures, temperature envelopes and plotting of geotherms - in other words to characterise the thermal regime at each site.

3.2 DATA BASE3.2.1 Ground Temperature Measurement

Ground temperatures were recorded at the eleven sites for periods ranging from 1 to 2.5 years by multisensor (thermistor) temperature cables, installed in boreholes. Borehole depths ranged from 10.7 m to 57.9 m and the number of sensors in any hole from 9 to 22. Sensors were more closely spaced near the top of the boreholes; spacing generally started at 0.75 m, increasing to 1.5 m, then to 3 m at the greatest depth. As many as 47 temperature logs were taken at a site over the observation period.

Resistance of the thermistors was measured using a modified Wheatstone bridge. Calibration of the sensors prior to installation enabled a rapid conversion from resistance to temperature. The temperature sensors, installation procedure, measuring equipment and measurement technique are

outlined in detail in Judge (1973a). The measurement system enables a resolution of 0.01 K, thus relative temperature changes in a particular sensor of 0.01 K can be detected although absolute temperatures are only accurate to 0.1 K.

3.2.2 Data File

The ground temperature data were available on a computer card file prepared according to the Earth Physics Branch (EPB) northern ground temperature data base. The EPB northern data handling programs were initially used to produce tables of all temperature logs, some plots of temperature versus depth, as well as temperature versus time at each site (see, for example, Taylor et al., 1982 for standard tables and plots). The preliminary data treatment revealed that, in addition to there being several erroneous values, probably resulting from errors made when transferring the data to card format, temperature readings were not recorded at regular intervals and gaps of up to 5 months occurred in several records. The data file was transferred to disk file on Carleton University's Honeywell CP-6 computer system for further handling.

3.3 PREPARATION FOR ANALYSIS

3.3.1 Quality Control

Before proceeding with a systematic quality check of the data files, the first few temperature logs, taken while ground temperature disturbance resulting from the drilling had not yet dissipated, were removed. A low resolution Fortran plotting program was developed to produce a separate plot of temperature versus time for each sensor at a site. Each plot was scaled based on the minimum and maximum temperatures recorded for the sensor; temperature variations at greater depth were thus more noticeable.

Erroneous points were removed from the data set by inspecting each plot of temperature versus time. Listings of temperature logs were also useful in the quality control and readily served to indicate if an entire temperature log was in error. For near surface sensors, where the annual ground temperature wave more closely reflects the annual air temperature wave, plots of air temperature from nearby weather stations were also helpful in the data inspection. Measurements where quality was uncertain, or where possible error was within the resolution of the measurement system, were not removed. Such points occurred particularly for deeper sensors, where the sinusoidal nature of the temperature wave was no longer perceptible, or changes of only a few hundredths of a degree occurred over the observation period, or both.

An additional component of data quality control consisted of ensuring that the minimum and maximum of each annual temperature wave was represented for each sensor. This step is important in later calculations of mean temperatures (particularly at shallower depths where temperature fluctuations are greatest). Minima and maxima had to be estimated in only a few cases.

A final component of the quality control, after data files were carefully edited, was the filling in of large (5 month) gaps in the temperature record. Calculations of mean annual ground temperatures and temperature envelopes for each sensor are greatly facilitated if, in addition to filling in such gaps, temperatures are interpolated at regular intervals. Several interpolating routines were tried before a satisfactory method was found. This procedure is discussed below.

3.3.2 Interpolating Routines

Three curve fitting and interpolating techniques were tested for their ability to best fit the measured temperature versus time data, and to predict temperatures over the 5 month gaps. The first approach consisted of a

Lagrange interpolating polynomial (Medearis, 1974, p.63-76). The Lagrange technique fits a unique polynomial to a set of unevenly spaced data points, the curve passing through all given points. The Lagrange technique was tested on the 0.75 m data at FGH-2 and NW-5. As Figures 3.1a and especially 3.1b show, the shape of the polynomial curve is not necessarily representative of the annual wave. The Lagrange interpolating polynomial technique was thus not pursued any further.

Two other techniques were then explored simultaneously; a cubic Spline interpolation routine and a Fourier series expansion. Each of these curve fitting techniques was used to interpolate temperatures at weekly intervals; and both the measured and predicted temperatures were plotted for each sensor.

The cubic Spline interpolation method (Medearis, 1974, p.67-71) fits successive polynomials to a set of data points, proceeding by groups of 3 points. At junction points, Spline produces a smooth transition, and at common points each polynomial gives the same value for ordinate and slope. The curve generated by the Spline routine passes through the given data points; however, it is not constrained by the minimum and maximum values and usually exceeds them.

Fourier's formula for complex harmonic functions was first applied to the study of annual temperature variations in the earth over a century ago (Thomson, 1860). The Fourier technique (e.g. see Carnahan et al., 1969, p.92) fits a fundamental sinusoidal curve to the data and a prescribed number of superimposed harmonics. A fundamental periodicity of one year was selected for the FGH - NW data, and a maximum of 20 harmonics was allowed in the series. This Fourier series curve does not necessarily pass through the measured data points and can either underpredict or overpredict the minima and maxima.

Figure 3.1 Lagrange interpolating polynomial fit to
0.8 m data at a) FGH-2 b) NW-5

Several manipulations of the data set were required before determining the Fourier coefficients.

- 1) The first and last temperature observations had to be separated by a period of time equal to the fundamental period, or some whole multiple thereof. The dates of the first and last logs were arbitrarily changed by up to two weeks to achieve this exact separation.
- 2) Observations at three of the FGH sites extended for 2.5 years. These data sets were therefore divided into 2 overlapping groups for the Fourier analysis: first 2 years, last 2 years.
- 3) To achieve the one year (or whole multiple) data separation for all sensors, occasional gaps in the first or last temperature logs were arbitrarily filled.

In addition if any linear trend (either of increasing or decreasing temperature) is superimposed on the annual wave, a correction must be applied to detrend the data before determining the Fourier coefficients. This correction can be added back on when generating data at weekly intervals.

Both the Spline and Fourier techniques seemed promising, the latter perhaps more so since the ground thermal regime is fundamentally harmonic and tends to repeat the same shape of curve from year to year. For deeper temperature sensors (greater than 10 m), where the variations are small and often masked by the measurement inaccuracies, the Spline curve bears no resemblance to an annual wave, whereas the Fourier curve is either sinusoidal or linear. The difference between the two techniques at these depths is perhaps not very important, because the annual temperature variations are so small (e.g. 0.15 K). For the shallower sensors, the ability of both techniques to fill in a 5-6 month gap was tested by removing observations from several sensors at a few sites and artificially creating a gap. Figures 3.2.,

B) FGH-3 4.6 m

Figure 3.2 Multiple harmonic Fourier series expansion and cubic Spline interpolation of FGH-3 data
a) 1.5 m b) 4.6 m

Figure 3.3 Multiple harmonic Fourier series expansion and cubic Spline interpolation of FGH-4 data
a) 1.5 m b) 3 m

Figure 3.4 Multiple harmonic Fourier series expansion and cubic Spline interpolation of 0.8 m data at FGH-4

3.3 and 3.4 show the temperatures predicted by Spline and Fourier over gaps created at FGH-3 (1.5 and 4.6 m), FGH-4 (1.5 and 3.0 m) and FGH-4 (0.75 m). The first two Figures suggest there is little difference in the techniques. However, Figure 3.4 reveals that, where the slope of measured temperature versus time changes markedly immediately after the start of a gap, Spline tends to maintain the previous slope resulting in a poor curve fit.

The Fourier expansion routine was therefore selected to interpolate over gaps in the observations, and to generate temperature data at weekly intervals. Additional subroutines were prepared to enable calculation of mean annual ground temperatures, annual temperature envelopes, to plot running mean temperatures (52 week periods), and to generate geotherm plots ready for temperature contouring. A typical output from a Fourier run for selected sensors at NW-7 is given in Appendix A1.

3.4 ANALYSIS: MEAN GROUND TEMPERATURES, ENVELOPES AND GEOTHERMS

3.4.1 Results

The Fourier program was run for each data set at each site, subdividing the data set into smaller time units where necessary. From the generated weekly data, mean annual temperatures and temperature envelopes were determined for each sensor and plotted for each site (Appendix A2). The NW data were readily separated into two years, 7/72 to 6/73 and 7/73 to 6/74. The FGH data, for comparison were divided into the same years with the initial six months incorporated into an overlapping first year 12/71 to 11/72.

The geotherm contour plots for each site are given in Appendix A3 and the running mean temperatures are plotted for each sensor at all sites in Appendix A4.

3.4.2 Discussion

Table 3.1 summarizes the basic information contained in the ground temperature profile data. The active layer (based on the Fourier predictions) ranges from 0.5 to 1.6 m and does not vary substantially over the observation period. Mackay (1981) noted that the presence of an ice-rich zone at the base of the active layer in Mackenzie Delta soils provides a large thermal mass which is insensitive to small changes in climate. The stable active layer depths at the study sites could therefore be due to the presence of an ice-rich zone at the base of the active layer; the visual ice logs (Isaacs, 1974) indicated that lenses are present at most locations.

Nearly all the mean annual temperature profiles show signs of a temperature decrease with depth, or a reduction in temperature gradient, suggesting an increase in surface temperature. Extrapolation of the temperature profile, at depths below the penetration of the annual temperature wave, to intersect with the surface, indicates possible changes in surface temperature of as much as 2.5 K.

Mean annual ground temperatures at a depth of 10.7 m vary from -0.94 to -2.3°C in 72/73, and in general increase from 72/73 to 73/74 by as much as 0.13 K. In fact, temperatures below the active layer show a clear increase over the observation period at all sites except NW-5, NW-6 and NW-7. At 0.8 m however, there is no consistent increase over the observation period: some sites underwent warming, others cooling. Temperatures at 0.8 m ranged from 0.83 to -2.25°C in 72/73 and from 0.09 to -2.33°C in 73/74.

The trends in near surface temperatures shown by the geotherm plots, in general indicate that the winter of 72/73 was warmer than in both 71/72 and 73/74, and that the summer of 73 was warmer than that of 72. This trend would seem to follow from the trends in air temperature and snowfall recorded at Norman Wells over the period, which are plotted in Figure 3.5 along with

TABLE 3.1

SUMMARY OF OBSERVED GROUND TEMPERATURE DATA

SITE	ACTIVE LAYER (m)	MEAN TEMPERATURE(10.7 m)		MEAN TEMPERATURE(0.8 m)	
		7/72-6/73	7/73-6/74	7/72-6/73	7/73-6/74
FGH-1	1.5	-1.00	-0.90	0.0	0.09
FGH-2	0.9	-1.33	-	-2.25	-
FGH-3	1.4	-1.68	-1.55	-0.60	-0.72
FGH-4	1.4	-1.33	-1.29	-0.51	-0.61
NW-1	1.6	-1.88	-1.84	-0.98	-2.16
NW-2	0.5	-2.30	-2.21	-1.65	-0.68
NW-3	1.1	-1.70	-1.60	-1.44	-1.90
NW-4	1.1	-2.29	-2.17	-1.41	-0.97
NW-5	1.3	-2.12	-2.10	-1.67	-1.68
NW-6	1.4	-0.94	-0.89	0.83	-0.64
NW-7	1.2	-0.97	-1.00	-1.87	-2.33

Figure 3.5 Air temperature and snowfall at Norman Wells, Sept. 1971 to Dec. 1974
(bars in snowfall plot indicate climatic normals)

climatic normal data. The plots of temperature versus time at 0.8 m for all sites (Figure 3.6), show that the trend varies from site to site and differences between sites vary from year to year. Figure 3.7 shows the running mean air temperature along with the running mean ground temperatures at 0.8 m; sites such as NW-1 and NW-6 follow closely the air temperature curve, but at others the resemblance is less marked, e.g. FGH-4 and NW-5. The running mean air temperatures are plotted starting from the period 1/72 to 12/72 and show a warming of ~ 1 K up to the period 7/72 to 6/73, when most of the ground temperature observations began. The running mean air temperatures seemed to stabilize until 1/73 - 12/73 and then decreased by ~ 0.8 K up to 7/73 - 6/74. The warming trend at the deeper sensors seen in the running mean plots in Appendix A3, could thus be a lagged response to the warming experienced at the surface in the several months prior to the observations. This lagged response is perhaps added on to a lagged response to the surface warming of as much as 2.5 K discussed earlier.

Chapter 5 analyses the interactions between ground temperatures and the surface climate and microclimate with the help of a numerical model, in an attempt to understand the reasons for the diversity of ground thermal conditions observed.

Figure 3.6 a) Plot of ground temperatures versus time at 0.8 m, Norman Wells sites

Figure 3.6 b) Plot of ground temperature versus time at 0.8 m, Fort Good Hope sites

Figure 3.7 Plot of running mean ground temperatures for all sites with a 0.8 m sensor and of running mean air temperature at Norman Wells

CHAPTER 4

SOIL PHYSICAL AND THERMAL PROPERTIES

4.1 INTRODUCTION

The ground thermal regime is a function of atmospheric climate conditions, the local microclimate, the geothermal heat from the interior of the earth and the thermal properties of the subsurface materials. The thermal properties of soil (conductivity, heat capacity and diffusivity) are, in turn, dependent on physical properties, including density, mineral constituents, grain size and water content. Some measurements of both physical and thermal properties were made on borehole samples from the study sites in the early 1970's, and this chapter presents the available data. Since a complete set of measurements was not performed on all soil layers, the bulk of this chapter therefore centers on methods to calculate physical and thermal properties from the available information at each site. Techniques to determine thermal diffusivity are explored at some length, as this property is of most interest in analysing transient heat conduction phenomena such as the annual cycle of shallow ground temperatures.

4.2 PHYSICAL PROPERTIES

In Chapter 2, stratigraphic logs were used to divide each site into simplified representative lithological units. Additional geotechnical data were also available from Judge (1973a), ~~Isaac~~ (1974) and Kurfurst (1973), and included gravimetric water contents, visual ice logs, dry bulk densities, specific gravity (particle density) and thermal conductivities (on frozen core samples). However, a complete set of measurements of thermal and physical

properties was generally not taken at each site for all soil units. Where several measurements of a property were taken within a lithological unit, the mean value was selected to represent the layer. The use of the series weighted average, which takes into account the depth interval over which a measurement applies, would have been more representative. However, information on depth intervals was lacking and this approach could not be used. Where no measurements of a property were made within a soil unit, a value from a similar lithologic unit at another of the study sites was adopted.

Calculations of porosity, volumetric water content and degree of saturation were made from the measured properties using the following formulas:

$$e = \frac{\rho_s}{\rho_d} - 1 \quad (4.1)$$

$$n = \frac{e}{1+e} \quad (4.2)$$

$$\theta = \frac{w \times \rho_d}{\rho_w} \quad (4.3)$$

$$S_r = \frac{\theta}{n} \quad (4.4)$$

where ρ_s = soil particle density

ρ_d = dry bulk density of soil

ρ_w = density of water

e = void ratio

w = gravimetric water content

n = porosity

θ = water content (volumetric)

S_r = saturation degree

Gravimetric water content was not available for all soil layers, but was estimated by assuming the soil to be saturated, and by solving the following equation for w :

$$\rho_d = \frac{1}{\frac{w+1}{\rho_s}} \quad (4.5)$$

Physical properties were compiled for all sites and are listed in Table 4.1 (which includes measured thermal conductivity, K , in the final column). The calculations indicated whether soil layers were unsaturated, saturated or whether they contained excess water (in the form of ice). Visual ice logs taken at several boreholes (Isaacs, 1974) indicated qualitatively whether a soil unit contained excess ice, e.g. by describing the soil as containing many ice lenses or as being an 'ice soil' mixture. At FGH-4 a description of 'ice-soil' agrees with the saturation ratio listed for the clay layer in Table 4.1. On the other hand, ice logs at NW-2 and NW 4 also suggested the presence of discrete ice lenses, yet the physical properties indicate the soil is unsaturated at depth. Such lack of corroboration may be because physical properties were measured only on the soil component of the soil-ice lens mixture and are thus not representative of the whole unit; alternatively, the soil component may be at a very low saturation level.

The volume fractions of soil solid components, i.e. the percentage of minerals and of organic matter, were calculated from the soil particle density, ρ_s . Values of mineral particle density of $2.6 - 2.7 \text{ Mg m}^{-3}$ and of organic particle density of $1.4 - 1.7 \text{ Mg m}^{-3}$ were assumed in the calculations (Andersland and Anderson, 1978). The determination of volume fractions (listed in Table 4.2) is useful to later calculations of thermal properties, which require a knowledge of the volume fractions of soil constituents.

TABLE 4.1. SOIL PHYSICAL PROPERTIES

SITE	LAYER	MATERIAL	REMARKS	w (%)	ρ_d (Mgm^{-3})	ρ_s	n (%)	θ (%)	Sr	K ($\text{Wm}^{-1}\text{K}^{-1}$)
FGH-1	0-.6 m	peat		159	0.27	1.65	85	43	0.5	
	0.6-1.5	sand		24	1.63	2.67	39	39	1.0	
	1.5-3.0	silt		26	1.58	2.68	41	41	1.0	1.55
	3.0-4.6	sand	excess ice	26	1.84	2.68	31	47	1.5	2.45
	4.6-13	silt	unsaturated	23	1.49	2.69	44	34	0.8	1.87
FGH-2	0-.5 m	peat		159	0.27	1.65	85	43	0.5	
	0.5-1.5	silty clay		80	0.80	2.10	62	62	1.0	
	1.5-6.4	till	unsaturated	9	2.00	2.70	25	17	0.7	1.10
	6.4-13	shale		8	2.30	2.75	20	18	0.9	1.20
FGH-3	0-1.7m	silty clay		40	1.25	2.48	50	50	1.0	
	1.7-6.4	gravel		21	1.72	2.72	37	37	1.0	2.40
	6.4-12.	clay		22	1.76	2.71	35	35	1.0	1.24
FGH-4	0-2.3m	silty sand		26	1.58	2.68	41	41	1.0	
	2.3-18	clay	excess ice	58	1.47	2.72	46	85	1.9	1.18
NW-1	0-3 m	sand		24	1.63	2.67	39	39	1.0	
	3-14	clay		36	1.36	2.66	49	49	1.0	1.53
NW-2	0-1.4m	silt/org	excess	145	0.65	2.10	69	94	1.4	1.49
	1.4-13.7	clay	unsaturated	35	0.95	2.70	64	33	0.5	1.46
NW-3	0-7.6	till	excess ice	30	1.91	2.61	37	57	1.5	
	7.6-6.1	shale		8	2.30	2.75	20	18	0.9	0.60
NW-4	0-.15	peat		150	0.56	2.00	70	95	1.4	
	.15-3	clay	unsaturated	63	0.86	2.61	67	54	0.8	
	3-14	clay	unsaturated	30	1.18	2.66	55	35	0.6	
NW-5	0-.15	peat		159	0.27	1.65	85	43	0.5	
	.15-3.2	till		25	1.59	2.65	40	40	1.0	
	3.2-6.7	shale	weathered	12	2.04	2.70	24	24	1.0	
	6.7-14	shale		9	2.17	2.70	19	19	1.0	
NW-6	0-3	peat	saturated	454	0.27	1.65	85	85	1.0	
	3-4.6	clay		46	1.16	2.50	53	53	1.0	
	4.6-10.4	siltstone		10	2.20	2.60	27	22	1.0	
	10.4-13.7	shale		8	2.20	2.72	19	19	1.0	
NW-7	0-1.2	clayey silt		80	0.80	2.10	62	62	1.0	
	1.2-4.6	sandy gravel		10	1.90	2.68	29	19	0.7	
	4.6-3.0	shale		8	2.20	2.72	19	19	1.0	

TABLE 4.2
SOIL SOLID COMPONENTS

<u>SITE/LAYER</u>	<u>% MINERAL</u>	<u>% ORGANIC</u>
FGH-1-1	-	100
1-2	100	-
1-3	100	-
1-4	100	-
1-5	100	-
FGH-2-1	-	100
2-2	50	50
2-3	100	-
2-4	100	-
FGH-3-1	85	15
3-2	100	-
3-3	100	-
FGH-4-1	100	-
4-2	100	-
NW-1-1	100	-
1-2	100	-
NW-2-1	85	15
2-2	100	-
NW-3-1	100	-
3-2	100	-
NW-4-1	-	100
4-2	100	-
4-3	100	-
NW-5-1	-	100
5-2	100	-
5-3	100	-
5-4	100	-
NW-6-1	-	100
6-2	90	10
6-3	100	-
6-4	100	-
NW-7-1	85	15
7-2	100	-
7-3	100	-

4.3. THERMAL PROPERTIES

4.3.1. Thermal Diffusivity

The shallow ground thermal regime is largely governed by the annual cycle of atmospheric conditions. The annual temperature wave is propagated into the ground with an exponentially decreasing amplitude and an increasing phase lag with depth. The lag and attenuation are functions of the thermal diffusivity of the substrate materials: diffusivity (κ) is the ratio of thermal conductivity (K) to heat capacity (C):

$$\kappa = \frac{K}{C} \quad (4.6)$$

The volumetric heat capacity, C , is the quantity of heat necessary to raise or lower the temperature of 1 m^3 of material by 1 K . The thermal conductivity ($\text{Wm}^{-1}\text{K}^{-1}$) of a material describes the rate of heat transfer by conduction through the substance subjected to a temperature gradient. more specifically, the quantity of heat that flows in unit time through a unit cross-sectional area and a unit thickness of material over which there is a unit temperature gradient. Thermal diffusivity (m^2s^{-1}) relates to the rate of change of temperature of a material subjected to a change of temperature at its surface. Diffusivity depends both on the ease with which heat flows in the substance (thermal conductivity) and on the heat capacity which governs the temperature change associated with the loss or gain of heat. The analysis of cyclic ground temperatures is the study of transient heat conduction. The differential equation for heat conduction (Fourier equation) relates the changing temperatures to the diffusivity as follows (for one dimensional flow):

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \quad (4.7)$$

In other words, the rate of change of ground temperature (T) with time (t) is a function of the thermal diffusivity, times the rate of change of the temperature gradient ($\partial T/\partial z$) with depth (z). The diffusivity is thus very important to the understanding of shallow ground thermal regimes. Other than calculating diffusivity from component properties (eq. 4.6), several methods exist for determining diffusivity directly: for example 1) by an analysis of the lag and/or attenuation of a temperature wave with depth 2) by solving equation 4.7 for κ (requires records of T versus t and z) and 3) by observations of crossovers of temperature profiles (Lovering and Goode, 1963).

In soils containing water, and subjected to temperature cycling above and below freezing, the diffusivity of the frozen soil will differ from that in the unfrozen state. This arises from the fact that the conductivity of ice is four times that of water, while the heat capacity of ice is roughly half that of water. Diffusivities of frozen soils should therefore be higher than the unfrozen values. In addition, soil moisture changes phase over a range of temperatures rather than at a particular temperature. This release or absorption of latent heat leads to an apparent heat capacity term and thus to an apparent diffusivity which is a function of temperature. Diffusivity thus changes as a function of depth and time in wet soil materials in permafrost environments. In using the ratio of heat capacity to thermal conductivity to calculate diffusivity, some of these factors may be overlooked. Direct determination of diffusivity from temperature observations at each depth may thus be advantageous.

4.3.1.1 The Fourier equation

The temperature data sets at NW and FGH lent themselves to an attempt to solve the Fourier equation (4.7) for diffusivity. Assuming heat flow to be one dimensional and conductive only, by evaluating the slope of the

temperature versus time curve ($\partial T/\partial t$) and the second derivative of the temperature-depth profile ($\partial^2 T/\partial z^2$) at time t and depth z , diffusivity can be determined for each depth at each site as a function of time (see Takagi, 1971; McGaw et al, 1978). This technique, requiring no laboratory measurements, could provide an elegant 'in situ' method of compiling a data set of diffusivities ideally suited to subsequent numerical modelling of ground thermal regimes.

The regularly spaced temperatures as a function of time generated by the Fourier expansion of Chapter 3, provided a data set suitable for the above calculations. A curve was fitted to the (T, t) data at each depth and the first derivative calculated at regular intervals of t . A curve was also fitted to the (T, z) data for each of the above t 's and the second derivative calculated at each z . The cubic Spline interpolation routine mentioned in Chapter 3 was readily applicable, as it allows for computations of first and second derivatives. The program was modified to handle data generated by the Fourier expansion and to calculate diffusivities bi-weekly.

A trial run of the program on portions of a data set where no major phase changes appeared to be occurring yielded ambiguous results. Some calculated diffusivity values were negative, i.e. not physically significant. To verify whether this was due to the numerical technique, a test data set was compiled of (T, z) for several depths, using the periodic equation of the surface temperature wave and its attenuation and lag with depth:

$$T_z = T_0 + A_0 \exp\left(-z \sqrt{\frac{\pi}{P \cdot \kappa}}\right) \sin\left(\frac{2\pi t}{P} - z \sqrt{\frac{\pi}{P \cdot \kappa}}\right) + g \cdot z \quad (4.8)$$

where A_0 = amplitude of surface temperature wave, T_0 = mean temperature at the surface, T_z = temperature at depth z , P = period of the wave, and g = geothermal gradient. Data were generated bi-weekly for a material with diffusivity $\kappa = 0.08 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$, $T_0 = 10^\circ \text{C}$, $A_0 = 8 \text{ K}$, $g = 20 \text{ mK m}^{-1}$ for depths of .1, .2, .3, ..., .9 m.

Calculated diffusivities gave positive values for .1, .2 and .3 m; however, at greater depths occasional negative values appeared. The negative diffusivities seemed to occur especially when the sign of the slope of temperature versus time changed. There may be inherent problems in the Spline approach where the rate of change of slope is very small; very closely spaced measurements may be necessary for the technique to work consistently.

Diffusivities calculated from the trial data set at .1, .2 and .3 m are plotted in a histogram in Figure 4.1. The values do cluster about a mean of $0.08 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$, the value used to generate the data set. Perhaps by ignoring the negative values calculated by Spline, a realistic estimate of diffusivity could still be obtained. A second data set was thus prepared, using the same parameters as before, but changing the depths to values similar to those at the study sites (0.75, 1.5, 2.3, 3.0, 4.6, 6.1, 7.6, and 9.1 m). A histogram showing only the positive diffusivities calculated for the 0.75 to 3.0 m depths yields a bi-modal distribution of values (Figure 4.2).

A final test of the method was undertaken on data sets from the study sites FGH-2 and NW 1. Negative diffusivities were ignored and a histogram of positive values plotted (Figure 4.3). The diffusivities show no tendency to cluster, but, rather, cover a broad range of values. In view of what was said earlier regarding changing diffusivity with changing temperature and phase of soil water, the scatter of values is not necessarily unreasonable. However, in terms of providing a set of diffusivity values versus time (such as FGH-2 in Table 4.3), there are large gaps where negative values have been ignored.

The negative values of diffusivity should perhaps not be dismissed so readily. The Fourier differential heat conduction equation (4.7) applies to one dimensional cases of heat transfer by conduction only. Both Lettau (1954) and Takagi (1971) have examined the relationship between $(\partial T / \partial t)$ and

Figure 4.1 Histogram of diffusivities calculated from the Fourier transient heat conduction equation and a test data set, for 0.1, 0.2 and 0.3 m

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Figure 4.2 Histogram of diffusivities calculated from the Fourier transient heat conduction equation and a test data set, for 0.75, 1.5, 2.3 and 3.0 m

Figure 4.3 Histogram of diffusivities calculated from the Fourier transient heat conduction equation and data from a) FGH-2, 1.5 - 6.1 m b) NW-1, 1.5 - 3 m and 4.6 - 7.6 m

TABLE 4.3

DIFFUSIVITIES CALCULATED FROM THE FOURIER EQUATION

AT FGH-2 (1.5 and 2.3 m)

Depth (m)	Time (days)	Temp (C)	$(\partial T / \partial t)$	$(\partial^2 T / \partial z^2)$	κ ($10^{-6} \text{m}^2 \text{s}^{-1}$)
1.5	7	-1.30	-0.084	-0.645	1.515
1.5	21	-2.62	-0.079	1.553	0.590
1.5	35	-3.38	-0.024	-0.532	0.525
1.5	49	-3.78	0.028	1.017	-0.327
1.5	63	-4.28	0.033	1.926	-0.198
1.5	77	-4.69	0.018	2.979	-0.068
1.5	91	-4.95	-0.010	3.764	-0.029
1.5	105	-4.95	0.020	3.783	0.062
1.5	119	-4.52	0.050	3.653	0.157
1.5	133	-3.81	0.061	3.261	0.217
1.5	147	-3.02	0.059	2.873	0.236
1.5	161	-2.35	-0.042	2.231	0.215
1.5	175	-1.91	0.027	1.324	0.238
1.5	189	-1.58	0.025	0.967	0.300
1.5	203	-1.32	0.019	1.091	0.200
1.5	217	-1.13	0.017	1.057	0.185
1.5	231	-0.94	0.016	1.074	0.169
1.5	245	-0.80	0.011	1.185	0.105
1.5	259	-0.70	0.011	1.045	0.118
1.5	273	-0.60	0.009	0.876	0.123
1.5	287	-0.54	0.007	0.711	0.001
1.5	301	-0.49	0.009	0.553	0.184
1.5	315	-0.46	0.003	0.517	0.060
1.5	329	-0.47	0.002	0.426	0.062
1.5	343	-0.46	0.006	0.418	0.154
1.5	357	-0.45	0.003	0.375	0.097
2.3	7	-1.00	-0.022	-0.979	0.258
2.3	21	-1.38	-0.026	-2.691	0.114
2.3	35	-1.69	-0.015	-3.911	0.046
2.3	49	-1.93	-0.018	-4.683	0.045
2.3	63	-2.20	-0.019	-5.399	0.042
2.3	77	-2.46	-0.015	-6.016	0.028
2.3	91	-2.67	-0.012	-6.345	0.022
2.3	105	-2.83	-0.005	-5.881	0.011
2.3	119	-2.83	0.007	-4.819	-0.018
2.3	133	-2.72	0.011	-3.293	-0.038
2.3	147	-2.58	0.011	-1.720	-0.078
2.3	161	-2.41	0.014	-0.359	-0.471
2.3	175	-2.19	0.018	0.337	0.624
2.3	189	-1.96	0.018	0.420	0.518
2.3	203	-1.73	0.016	-0.008	-22.671
2.3	217	-1.53	0.015	-0.214	-0.808

$(\partial^2 T / \partial z^2)$, the former for near surface soils, the latter for a lake.

Lettau's work indicates that the relationship between the two derivatives is not linear but may be elliptical (Figure 4.4) leading to negative values of $(\partial^2 T / \partial z^2)$ for positive values of $(\partial T / \partial t)$, or vice versa, and thus to negative diffusivities calculated from the Fourier equation. Lettau attributes this to processes other than simple heat conduction occurring, e.g. moisture flows and other heat transfer processes, as well as to the fact that diffusivity is depth and time dependent. Takagi (1971) applied the Fourier equation to observations of lake temperatures (where the thermal diffusivity was known) and calculated the residuals, $(\partial T / \partial t) - \kappa \cdot (\partial^2 T / \partial z^2)$, to analyse how the actual heat transfer mechanisms might have differed from simple unidirectional heat conduction.

The negative diffusivities calculated from the Spline program may therefore reflect, in part, problems with the Spline function due to the data spacing and, in part, complex heat transfer processes occurring at the sites. Clearly the technique cannot be used to systematically generate diffusivities.

4.3.1.2 Attenuation and lag of surface temperature wave

The classic method of calculating diffusivity from the attenuation and lag of the surface temperature wave was next examined. This method uses the ground temperature data base to calculate the attenuation of the annual wave, and/or the phase lag between any two depths, to solve for the diffusivity over the depth interval, using equation (4.8). The classical heat conduction model is again assumed in this analysis.

From the ratio of the attenuation of the amplitude between two depths, z_1 and z_2 , the following relationship is derived from (4.8):

$$\kappa = \frac{\pi}{P} \left[\frac{z_2 - z_1}{\ln(A_{z_1} / A_{z_2})} \right]^2 \quad (4.9)$$

Figure 4.4 Relations between partial derivatives of soil temperatures (Lettau, 1954):
 a) schematic b) observed diurnal, 2-5 cm
 I = linear, II = elliptical

and from the ratio of phase lags:

$$\kappa = \frac{P}{\pi} \left[\frac{z_2 - z_1}{t_2 - t_1} \right]^2 \quad (4.10)$$

This technique was used long ago by Thomson (1860) who combined the diffusivities with measurements of heat capacity to derive thermal conductivity. The calculations for the NW and FGH sites were performed on the data synthesized from the Fourier expansions. Diffusivities could not be calculated for every soil unit because 1) there was not always two temperature sensors within a layer, or 2) at greater depths the attenuation of the wave was such that minima and maxima, and times of occurrence could not be discerned, or 3) the temperature changes became linear at depth or 4) the same peak and trough could not be followed through from top to bottom in a profile.

The calculation of diffusivity from the phase lag was hampered by difficulties in defining the exact time of occurrence of minima and maxima. Results seemed erratic and often an order of magnitude greater than values derived from the attenuation. The phase lag method was thus rejected in favour of the attenuation. Such order of magnitude discrepancies have been reported elsewhere (e.e. Lettau, 1954) and may not be due solely to computational inaccuracies.

In order to ensure that the best possible estimate of diffusivity was obtained from the attenuation, the possibility that the temperature record may have missed the maximum and minimum near the surface, thus underestimating the attenuation and overestimating the diffusivity, was examined. The temperature logs at NW and FGH were taken approximately 25 times in a year. Twenty five numbers were randomly generated from 1 to 365 and temperatures calculated for

these 25 days for ten depths using the periodic equation (4.8). Once again $T_0 = 10^{\circ}\text{C}$, $A_0 = 8\text{K}$, $\kappa = .08 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ and $g = 20 \text{ mKm}^{-1}$ and depths ranged from 0, .1, .2 to .9 m. The amplitude of the wave yielded by random selection was then compared to the real amplitude. This process was repeated for 10 different series of randomly generated numbers. The average of 100 comparisons gave an amplitude derived from the 25 points which underestimates the real amplitude by an average of 1.1%; the underestimation ranged from 0 to 5.6%.

The temperature logs at NW and FGH were, however, not taken completely at random, but often bi-weekly. Thus a second comparison was made, taking 25 days spread at two week intervals and repeating this ten times, each time starting with a different day in the year (i.e. from day 9 to day 18 as starting points). This exercise indicated that the amplitude would always be underestimated by less than 1%. Thus if the temperature wave were perfectly sinusoidal, then the data sets at NW and FGH should only underestimate the amplitude at any given depth by about 1%.

The actual irregularities in the surface temperature wave, and the several months gap in the data, no doubt increase the underestimate of the amplitude. This is most likely to occur for near surface sensors where the irregularities in the annual wave are not yet dampened out. A greater inaccuracy in estimating the amplitude at these depths might arise, however, from the use of data generated by the Fourier expansion. The curve from the 20 harmonic expansion did not fit perfectly the irregularities in the near surface temperature data and residuals (observed minus predicted values from the Fourier expansion) of as much as $\pm 0.6\text{K}$ occurred at depths of 0 and 0.75 m.

Amplitudes were therefore determined from measured data at a few sites and compared to values derived from the synthesized data. The synthesized amplitudes ranged from 10% more to 10% less than the measured; in the worst case this could lead to an error of almost an order of magnitude in the calculation of diffusivity.

These analyses suggest that the use of the data generated by the multiple harmonic Fourier expansion can lead to greater errors in estimating the amplitude of the annual wave in the ground than can the irregular spacing of the data logs. Diffusivities calculated from the attenuation of the amplitude, using measured data and synthesized data, are listed in Table B1 in Appendix B. Differences of over 200% can be seen; the synthesized data set is thus not suitable for these calculations.

4.3.1.3. Discussion

The attenuation technique is not applicable to every soil unit, even if the measured data set is used for calculations. As Table B1 indicates, quite different values of diffusivity can be obtained for a soil layer depending on which pair of sensors is selected, or which year (i.e. which temperature wave) is analysed. This is due in part to inhomogeneity within a soil unit. Where values appear to be within the right range (compared, for example, to typical published values shown in Table 4.4), they might represent an average over a broad range of temperatures (passing from positive to negative, with the soil undergoing phase change) and thus an average of unfrozen, freezing and 'frozen' diffusivities. Very few diffusivities can be seen to be representative of a narrow range of temperatures, and consistent within a layer.

To summarise, the temperature data do not appear to be able to yield thermal diffusivities either as a function of depth, temperature or time. Lovering and Goode (1963) also developed methods for calculating thermal

TABLE 4.4. PUBLISHED VALUES OF THERMAL DIFFUSIVITY

<u>MATERIAL</u>	<u>REMARKS</u>	<u>THERMAL DIFFUSIVITY</u> ($\text{m}^2\text{s}^{-1} \times 10^{-6}$)
sandy soil	dry	.24
	saturated	.74
clay soil	dry	.18
	saturated	.51
peat soil	dry	.10
	saturated	.12
ice		1.16
water		.14
gravel		.57-.62
shale		.40
sandstone		1.20-1.40

Sources: Oke (1978) and Lovering and Goode (1963)

diffusivity from borehole temperature measurements. Their techniques combine graphical and mathematical methods and use either of two relations: 1) that the slope reversal points on temperature depth curves (i.e. where the slope changes sign) are separated by half a wavelength or 2) that the vertical separation of successive crossings of any two curves is always equal to one half the wavelength. The techniques are, in fact, difficult to apply because of inaccuracies in determining when crossovers occur, or when slopes change sign - particularly because of the small range in temperature at depth and the presence of a geothermal gradient. As well, these techniques apply to a homogeneous medium, not to layered soils such as those at the study sites. Another of Lovering and Goode's methods requires two temperature profiles, a knowledge of their crossover, depth and of the time in the spring when the annual temperature wave passes through the average surface temperature. This last prerequisite is not easily determined at the FGH and NW sites, because of the paucity of data at the surface. Thus, none of Lovering and Goode's techniques were attempted in the present study, due to the coarse data spacing in depth and time.

Despite the fact that diffusivities cannot be determined from the borehole temperature data, they still may be calculated from heat capacities and thermal conductivities. The next section examines the use of soil physical properties and measured thermal conductivities to determine heat capacity and thermal conductivity for each soil unit over a range of temperatures.

4.3.2. Heat Capacity and Thermal Conductivity

4.3.2.1. Heat capacity

The heat capacity (C) of a unit volume of soil can be determined from the sum of the heat capacities of the soil constituents. Where no phase transformation occurs, the heat capacity is given by:

$$C = C_s X_s + C_i X_i + C_a X_a + C_u X_u \quad (4.11)$$

$$\text{where } C_s = C_m X_m + C_o X_o$$

C_n = heat capacity of constituent n

X_n = volume fraction of constituent n

n = s (solids), m (minerals), i (ice), o (organic),
u (water), a (air)

Heat capacities for soil constituents are given by de Vries (1975):

$$C_m = 2.0, C_o = 2.5, C_u = 4.2, C_i = 1.9, C_a = .00126 \text{ (MJm}^{-3}\text{K}^{-1}\text{)}.$$

Volume fractions of soil constituents for the unfrozen state (temperatures greater than or equal to 0°C) were calculated from the physical properties, and the heat capacity at positive temperatures was calculated for all layers. Heat capacities at several negative temperatures (-0.5° , -1.0° , -2.0°C) were also determined, with the value at -2.0°C assumed to represent colder temperatures (since, for most soils, very little change occurs in the unfrozen water content at temperatures below -2°C). In order to determine the volume fractions of unfrozen water and ice (not taking into account the volume expansion of water when it changes to ice), representative values were selected from freezing characteristic curves of major soil types published in Smith and Patterson (1982), Farouki (1981) and Nakano and Brown (1972). Some typical curves are shown in Figure 4.5.

4.3.2.2. Thermal conductivity

Thermal conductivity measurements made on selected samples from the NW and FGH sites are reported by Judge (1973a). The measurements were made in a

Figure 4.5 Soil freezing characteristic curves (after Smith and Patterson, 1982)

cold room using the divided bar technique, which is described in detail by Judge. The thermal conductivities were determined on frozen core samples at temperatures of -2.5°C or colder, to ensure that most of the water was frozen, and that only conductive heat transfer processes occurred during measurement. Measurements were not taken on unfrozen samples, nor over the temperature range of 0° to -2.5°C , the range of temperatures actually encountered in most of the NW and FGH boreholes. Over this range, the conductivity can change dramatically in response to changing ice/water ratios. Thermal conductivities thus had to be calculated for soil units where no divided bar measurements were available, and for soil temperatures greater than -2.5°C .

Farouki (1981), in a comprehensive review, compares many methods for calculating soil thermal conductivity (e.g. Kersten, Johansen, de Vries). For the soil types and degree of saturation encountered at the study sites, Farouki recommends Johansen's formulas (which are based on the geometric mean method) for ease of application as well as quality of results. For saturated soils, these formulas are:

$$\text{i) } K_{\text{sat}} = K_s^{(1-n)} K_{\theta}^n \quad \text{for unfrozen soil} \quad (4.12)$$

$$\text{ii) } K_{\text{sat}} = K_s^{(1-n)} K_i^{(n-\theta)} K_{\theta}^{\theta} \quad \text{for frozen soil} \quad (4.13)$$

where K_s = soil solids conductivity

K_{θ} = conductivity of water = $0.57 \text{ Wm}^{-1}\text{K}^{-1}$

K_i = conductivity of ice = $2.2 \text{ Wm}^{-1}\text{K}^{-1}$

θ = unfrozen water, volumetric content

n = porosity

and for unsaturated soils:

$$K = (K_{\text{sat}} - K_{\text{dry}}) K_e + K_{\text{dry}} \quad (4.14)$$

where $K_e = 0.7 \log S_r + 1.0$ for coarse unfrozen soils

$K_e = \log S_r + 1.0$ for fine unfrozen soils

$$K_{dry} = \frac{0.135 + 64.7}{2700 + 0.94\rho_d}$$

ρ_d = dry bulk density

S_r = saturation ratio

Farouki points out, however, that one difficulty with Johansen's method is in determining the soil solids' conductivity. Raj and Judge (unpublished manuscript) determined values ranging from 3.2 to 9.0 $\text{Wm}^{-1}\text{K}^{-1}$ for sand solids, 2.4 to 5.9 $\text{Wm}^{-1}\text{K}^{-1}$ for silt solids, and 2.8 to 4.8 $\text{Wm}^{-1}\text{K}^{-1}$ for clay solids. Soil solid conductivity was calculated for each soil unit at the study sites using the median values of 6.1, 4.2 and 3.8 $\text{Wm}^{-1}\text{K}^{-1}$ for sand, silt and clay respectively, and a value of 0.3 for peat solids and 1.5 for shales. Particle size analysis information on the core material (unpublished data provided by P. Kurfurst, Geological Survey of Canada) was consulted to verify lithological descriptions and determine the percentage of sand, silt and clay size fractions in a soil unit. Freezing characteristic curves provide data on unfrozen water content as a function of negative temperatures and enable the calculation of thermal conductivities as a function of freezing temperature.

Where results of divided bar measurements on frozen soil samples were available, Johansen's formulas could be solved for K_s , and this value used in turn to calculate the soil thermal conductivity at 0°, -0.5° and -1.0°C.

4.3.2.3, Results

Calculated heat capacities and calculated and measured thermal conductivities are listed in Table B2 in Appendix B. A comparison of the

measured and calculated conductivities reveals that the latter are higher, by about 60% on average. Whether this difference can be attributed to errors in estimating the volume fractions of soil constituents, or to incorrect values assumed for soil solids conductivity, or to both, cannot be determined.

However, where the Johansen calculated values are the only ones available for a unit, they have been arbitrarily reduced by 50% for use in the numerical modelling in Chapter 5.

Table B2 also lists thermal diffusivities calculated from the heat capacities and the measured thermal conductivities, or the reduced calculated values. The heat capacities and thermal conductivities were input into the numerical model (Chapter 5) and used to generate diffusivities. The modelling more specifically requires a freezing characteristic curve for each soil unit, and thermal conductivities as a function of negative temperature (3-4 data points were sufficient to define each curve). When phase change occurs, an apparent heat capacity term is determined for each temperature from:

$$C_{app} = C + L\rho_w \left(\frac{d\theta_{uf}}{dT} \right)_T \quad (4.15)$$

C_{app} = apparent heat capacity

C = heat capacity of soil constituents (see equation 4.11)

$L\rho_w \left(\frac{d\theta_{uf}}{dT} \right)_T$ = heat absorbed or released by phase change at temperature T

L = latent heat of freezing

ρ_w = density of water

$\left(\frac{d\theta_{uf}}{dT} \right)_T$ = slope of the freezing characteristic curve at temperature T

θ_{uf} = unfrozen water (volumetric)

The thermal conductivity at any temperature is determined by interpolation of the conductivity-temperature data.

Several sites have a surficial layer of peat. However, only at NW-6, where this layer is 3 m thick, were physical properties available for calculations of thermal properties. At the remaining sites, bulk densities were only rarely measured, and water contents and degree of saturation could only be estimated from the general site description. Thus calculations of thermal properties for peat layers can only be very approximate, especially given the wide ranges of density and gravimetric water contents encountered in peats (for example see Farouki, 1981). The lack of specific information is all the more unfortunate, because the surficial layers are most important in understanding the ground thermal regime.

4.3.3 Summary

The thermal property of most interest in analysing transient heat transfer, such as the annual cycle of shallow ground temperatures, is the thermal diffusivity. When laboratory measurements are not possible or available, thermal diffusivity may be determined by calculations based on the soil physical properties, or on a series of ground temperature observations. Numerical techniques relying on the use of ground temperature data to determine diffusivity values by either 1) solving the Fourier transient heat conduction equation for diffusivity (eqn 4.7), or 2) analysing the phase lag and amplitude attenuation of the annual surface temperature wave (equations 4.9 and 4.10), cannot be systematically employed with confidence for the soil and temperature conditions at NW and FGH. The calculations are hampered by several factors: 1) the irregular spacing of temperature measurements both in time and with depth, 2) soil inhomogeneity, and the lack of more than one temperature sensor within each soil layer, 3) heat transfer processes other

than pure conduction - such as heat transfer due to moisture flow, and 4) latent heat effects which give rise to temperature dependent diffusivities.

In the absence of laboratory measurements, thermal diffusivity must be determined from the ratio of thermal conductivity and heat capacity. These properties may be calculated from soil physical properties. Heat capacity can be calculated from a knowledge of the volume fractions of soil constituents and of the soil freezing characteristic curve. Thermal conductivity computations also require a freezing characteristic curve; if none has been measured then the grain size characteristics and the mineralogical composition of the soil are needed to select a representative curve from the literature. Johansen's formula may then be used to calculate conductivity, although it is not readily applied unless soil solids thermal conductivity is known. Again, a knowledge of grain size and mineralogy will help to select a value from the sparse literature.

In the present research, thermal conductivity values derived by Johansen's method were 60% higher on average than measured conductivities. This discrepancy may be due to errors in the selection of K_g , to errors in the degree of saturation, to errors in the selection of a soil freezing characteristic curve, or to some combination of the above. The effect on the ground thermal regime of such errors in thermal conductivity is examined in Chapter 5, by varying the thermal conductivities by $\pm 50\%$ and simulating ground temperatures. Some estimate of the percentage error that may be tolerated in thermal conductivity values can then be made, depending on the desired application.

CHAPTER 5

NUMERICAL MODELLING OF MICROCLIMATIC AND GROUND THERMAL REGIMES

5.1. INTRODUCTION

"The objective of ... a modeling effort is not merely a presentation of field observations in mathematical language, but validation of physical laws upon which the model is based and determination of the relative importance of associated physical parameters. The results of mathematical modeling and validation also guide field observers in designing additional studies."

Nakano and Brown (1972, p.20)

In Chapter 2, the factors influencing the ground thermal regime were presented in some detail. Temperature conditions within the ground are bounded, on the one hand, by the surface temperature, and, on the other, by the temperature at depth. Temperature increases with depth due to the geothermal gradient, and the rate at which temperature increases is a function of the geothermal heat flux and of the thermal properties (and hence lithology) of the substrate. The surface temperature is governed by the surface energy exchange which is driven by solar radiation. The energy balance is in turn a function of the local climate and the microclimatic setting. Surface temperatures vary daily and seasonally; mean annual surface temperatures also vary on a longer term, in response to large scale variations in climate. Changes in surface temperature may arise as well from natural or man-induced changes to the local ground surface conditions, e.g., forest fires or highway construction. The propagation into the ground of these surface

temperature changes depends on the thermal diffusivity of the earth materials, which, in permafrost soils is very much dependent on the presence of soil moisture and on the release or absorption of latent heat.

This chapter examines the interaction of climate, microclimate and ground thermal regimes at the study sites via a computer simulation model. The modelling was undertaken in three steps, to determine: 1) the influence of the observed lithology on the ground thermal regime, 2) the effect of natural variations in climate occurring over the observation period, and 3) the role of microclimatic variations at the sites. In addition, results of the analysis can be used to determine the important physical parameters for measurement in future studies of shallow ground thermal regimes. Preparation of the data set for the computing runs, and application of the model to the field conditions, required several refinements to the computer program. This procedure is presented in detail as it serves to stress the limitations in both the background data available for input, and in the ability of simulations to accurately reproduce field conditions.

5.2. THE MODEL

The computer simulation of microclimatic and ground thermal regimes described by Smith (1977) was used to simulate the relationship of climate and ground thermal regimes at the study sites. The program requires input of weather data and site specific characteristics in order to constrain the surface energy balance:

$$Q^* + Q_H + Q_E + Q_S = 0 \quad (5.1)$$

where Q^* = net radiation balance, Q_S = conduction of heat in the subsurface, Q_E = transfer of heat by evaporation (condensation), and Q_H = convective transfer of heat into the air.

Qualitatively each component can be expressed in functional relation as follows:

$$Q^* = f(\text{albedo of surface, emissivity, air temperature, cloud cover, cloud type, surface temperature, solar radiation})$$

$$Q_E = f(\text{wind speed, air temperature, surface temperature, air and surface relative humidities, aerodynamic roughness})$$

$$Q_H = f(\text{wind speed, aerodynamic roughness, surface temperature, air temperature})$$

$$Q_S = f(\text{thermal conductivity of surface layer of soil, surface temperature, temperature at depth})$$

The above parameters, except the surface temperature, are input or derived from input data. The value of surface temperature which solves the energy balance equation, the equilibrium surface temperature, is arrived at by an iterative technique. The surface temperature is then used as a boundary condition for the soil thermal diffusion and, in conjunction with the thermal properties, to generate a ground temperature profile. A one-dimensional implicit finite difference solution of the heat conduction equation is used. The principle of the operation of the model is outlined in Figure 5.1.

The Fortran program (hereafter called MICRO) calculates the surface energy balance on a daily basis, using either daily weather input, or interpolated daily weather from mean monthly values. Site specific characteristics (albedo, roughness and wetness), either from field measurements or taken from the literature, are input at the start of the program and modified internally as a function of snowcover conditions. An initial ground temperature profile is specified. Any number of soil layers can be accommodated; thermal properties for frozen and unfrozen states must be specified. The model has recently been modified (Smith and Riseborough, pers. comm.) to allow thermal properties to vary fully as functions of negative temperature - a knowledge of the freezing characteristic of the soil and of

Figure 5.1 Flow chart of microclimatic and ground thermal regimes program; Q^* : net radiation, Q_E : evaporative heat flux, Q_H : convective heat transfer into air, and Q_S : soil heat flux (Smith, 1977)

the thermal conductivity as a function of temperature is required. Latent heat energetics are incorporated into the apparent heat capacity (see section 4.3.2.3.). The new version also allows for daily inputs of snowfall. The node spacing of the finite difference grid is set to 5 cm in the snowpack and in the upper 50 cm of the ground; the spacing then increases exponentially down to 100 m.

MICRO has been tested and found to perform satisfactorily in prediction of ground temperatures in permafrost and non-permafrost situations. Active layer development in the high Arctic was modelled by Smith (1977) and frost penetration beneath highways by Smith and Tvede (1977). During the course of the present research several difficulties were encountered; refinements either to the model or to the way in which it was used were required. These are outlined in the following sections.

5.3. PROBLEMS WITH THE ANALYSIS

To determine the effect of site lithology on the ground thermal regime, simulations imposing the same microclimatic conditions and atmospheric conditions on all sites were run. The microclimatic conditions were selected to be representative of a non-forested site with a surface albedo of .20, aerodynamic roughness of 5.0 cm, and a surface wetness of 0.5. Albedo and roughness values were selected from the literature (Oke, 1978; Burns, 1973); surface wetness was assigned based on the average drainage conditions. All sites were assumed to be on flat surfaces at the same altitude. Mean monthly climatic normals (30 year average) for Norman Wells were used for air temperature, precipitation, solar radiation, wind speed, relative humidity, cloud cover, and pressure (Phillips and Aston, 1980; Environment Canada, 1973a, 1973b and 1975; Ministry of Transport, 1968a and 1968b). Information

on average monthly cloud type was found in Vowinckel (1962), and for snow densities in Burns (1973). To generate an initial ground temperature profile, a mean surface temperature (T_s) of -2.5°C was chosen, based on the extrapolation of temperature gradients at the deeper sites, and a geothermal gradient of 60 mKm^{-1} was assumed based on local data (Judge, 1973a).

Simulations were set up for each site with respective subsurface lithologies, and the physical and thermal properties established in Chapter 4. The MICRO runs were started in July and run for 5 years to ensure that the ground thermal regime achieved an equilibrium with the surface and climatic conditions. Several difficulties became apparent as the first simulations were performed.

5.3.1 Disequilibrium Profiles

Figure 5.2 shows the results of an early run for FGH-4. The mean annual temperature profile after the first complete calendar year is colder than the initial profile; the second year is much warmer and closer to the starting conditions. Slight warming continues over the next 3 years. The simulation begins in July but with 'mean' annual ground temperatures, whereas near-surface temperatures in July would actually be warmer. The simulation thus produces unrealistically low temperatures at the outset. After a few years the mean annual temperature at the surface seems to have reached an equilibrium value, although slight warming at depth is still noted.

A longer simulation (Figure 5.3) shows that ground temperatures continue to increase from 5 to 10 years after the start of the run. Upon examination the equilibrium surface temperature after the first few years was found to be 0.5°C warmer than the initially prescribed temperature. Thus, in effect,

Figure 5.2 Mean ground temperature profiles and envelopes for 0, 1, 2, and 5 year numerical simulations at FGH-4 with climatic normals and initial $T_s = -2.50C$

Figure 5.3 Mean temperature profiles and envelopes for 5 and 10 year simulations at FGH-4 with climatic normals and initial $T_s = -2.5$ °C

the warming of the ground is in response to a change of surface temperature. Calculations based on the method outlined by Birch (1948) to determine the disturbance to temperature in the earth resulting from climatic fluctuations, indicated that the changes observed from 5 to 10 years after a surface disturbance of 0.5 K are of the right order of magnitude. The initial selection of the ground temperature profile can thus lead to disequilibrium conditions, particularly if a simulation is run only for one year, or if the initially prescribed surface temperature (T_s) differs from the equilibrium value calculated by the model.

To overcome these difficulties, the following procedure was devised:

- i) start with an estimate of T_s and g (-2.5°C and 60 mKm^{-1} in this case) and run the simulation with climatic normals for 2 years to get an equilibrium T_s .
- ii) take the new T_s and once again add the geothermal gradient and run with the climatic normals for 5 years. (The first years's mean annual profile will still be disturbed by the fact the simulation starts in July, and slight warming might still be occurring at depth over 5 years.)

5.3.2 Snowdepths

Goodrich (1982) comments: "...because it is difficult to model snowcover accurately the reliability of ground thermal regime computations is adversely affected". The influence of snowcover on the ground thermal regime in permafrost areas is discussed qualitatively at some length by Goodrich. He shows that models of the effect of snowcover on mean annual ground temperatures in permafrost regions are extremely sensitive to assumptions made in treating the snow cover - i.e. how it is allowed to build up, total thickness, and duration. MICRO was modified to record snowfalls of as little as 0.1 cm, to help model the snowpack more realistically. The program was originally sensitive to snowfalls of 0.5 cm or more.

In the present version of MICRO, snowfalls are input on a daily basis. The density of freshly fallen snow is less than that of the established snowpack. The evolution of the snowpack is a complex phenomenon (see Gray and Male 1981, chapters 5 and 7). Many factors are important, e.g. the weather (temperature, wind and moisture both during and after precipitation), the processes of metamorphism and ablation, topography and vegetative cover. Systems for recording snowfall also vary. At 85% of the AES stations, snowfall measurements are done with a ruler, and for this type of recording the time of measurement is a crucial factor affecting the thickness. There is no simple formula for converting depth of freshly fallen snow into thickness of snowcover.

By simply inputting 100% of the snowfall into the model, a thicker snowcover accumulates than is recorded on the ground by AES. One way to overcome the problem is to reduce snowfall before input into the simulation. This reduction can be determined by trial and error. Snowdepths predicted by MICRO using 66% of the climatic normals snowfalls agreed reasonably well with those recorded by AES (Figure 5.4).

Snowfall reduction factors were also determined for the winters of 72/73 and 73/74 to use in the second phase of the simulations. Figure 5.5 shows that these two winters differed from the normal and that inputs of 66% of snowfall did not produce MICRO snowdepths which agreed with those recorded by AES at Norman Wells. Reductions of one quarter in 72/73 and of one half in 73/74 gave more acceptable results. The incorporation of a model of snow accumulation and ablation would have been a more elegant solution to the problem. However, the development of such a model was not a focus of the research; the concern was rather to ensure that the snowdepth conditions were realistically reproduced. The arbitrary reduction of snowfalls achieved this with the most expedience.

Figure 5.4 Climatic normal snowdepths recorded by AES at Norman Wells, and predicted by MICRO for various snowfall inputs

Figure 5.5 Snowdepths recorded by AES Norman Wells and predicted by MICRO for various snowfall inputs a) 1972-1973 b) 1973-1974

Figures 5.4 and 5.5 indicate that MICRO predicts well the evolution of the snowpack in the early and mid-winter, but not in spring. In the predictions, ablation begins later and, as a result, greater snowdepths persist into the spring. The ablation period is also shorter with the snow cover disappearing in about a week. The ablation of the actual snowpack occurs over a longer period - although the exact length of time is difficult to judge on the basis of 'end of the month' recorded snowdepths only. Snowdepth probably starts to decrease later in April, and not at the beginning (as shown in Figures 5.4 and 5.5).

While determining the reduction factor for 73/74, a simplification resulting from the use of monthly weather data was underlined. In January 1974, AES recorded little change in snow on the ground at the end of the month, compared to the observation for December 1973; yet, 25.7 mm W.E. of precipitation fell in January. A detailed inspection of the daily temperature records revealed that during a week in early January, the maximum daily temperature had been above freezing. Not only did ablation of the snowpack probably occur at this time, but some of the precipitation may also have been in the form of rain. The use of monthly mean temperatures interpolated to obtain daily temperatures naturally overlooks such short term fluctuations in weather.

5.3.1.3. Large Temperature Inversions

A final difficulty with the simulations was noted upon closely examining the energy balance components. In Chapter 2 the occurrence of the Arctic Inversion (air temperature increasing with height near the ground due to radiative cooling) was noted. Burns (1973) reported the frequency of inversions at Norman Wells throughout the year (Table 5.1); in mid-winter,

TABLE 5.1.

ARCTIC INVERSIONS AT NORMAN WELLS

PERIOD	% OF TIME INVERSION PRESENT	
	at night	late afternoon
Sept-Nov	49	22
Dec-Feb	59	49
March-May	65	13
June-Aug	69	5

22
OF / DE

they occur 50% of the time. Weller and Holmgren (1974) show temperature profiles over the arctic tundra in winter with steep and stable inversions of about 4 K between the snow surface and 2 m above (typically, 13 K inversions occur between the snow surface and 16 m). Geiger (1975, p.215) published values of night inversions measured over snow: for wind speeds of 3 to 6 ms^{-1} an inversion of roughly 5 K was observed between 1 m and 1.4 m. Air temperatures used in MICRO correspond to the 2 m height of standard observing stations. Inversions of 1 or 2 K seem more realistic for the simulations, in light of the fact that the model represents an average of daily, nightly, and monthly variation. Yet, winter inversions persisted for some months at magnitudes between 5 and 10 K (Table 5.2, column 2). The large inversions in 72/73 and their absence in 73/74, led to mean annual ground surface temperatures in 73/74 that were warmer than those of 72/73, contrary to what might be anticipated from the records of air temperature and snowfall for these years.

To avoid spurious occurrences of large temperature inversions, the program was modified by imposing tighter constraints on the possible range of values for the equilibrium surface temperature and on the accuracy with which the energy exchange terms balance. The inversions predicted with these modifications are listed in the final column of Table 5.2, and, with one exception of 4 K, are on the order of 1 K.

5.4 SIMULATIONS

5.4.1 Lithology Modelling

Simulation runs were prepared for each site with its respective subsurface conditions, but imposing identical surface conditions (as described in section 5.3) and climatic normals from Norman Wells across all sites. The

TABLE 5.2.

INVERSIONS PREDICTED BY MICRO

	<u>Tair</u>	<u>Inversion</u>	
		original	modified MICRO
Normals data			
Oct	-4.1	1.3	1.3
Nov	-18.1	9.4	1.0
Dec	-25.9	6.6	0.8
Jan	-28.7	0.9	0.9
Feb	-26.6	0.8	0.8
Mar	-18.8	1.0	1.0
72/73 data			
Oct	-4.0	1.1	1.1
Nov	-18.0	10.2	4.0
Dec	-23.8	7.8	0.9
Jan	-28.2	5.8	0.8
Feb	-24.2	6.7	0.9
Mar	-21.7	0.6	0.6
73/74 data			
Oct	-4.6	0.9	0.9
Nov	-20.6	1.2	1.2
Dec	-23.9	1.0	1.0
Jan	-24.9	1.1	1.1
Feb	-30.6	0.7	0.7
Mar	-22.3	0.9	0.9

differences in ground temperature profiles resulting from this analysis should be related solely to the variations in subsurface lithology. The MICRO runs were performed in two steps as outlined in section 5.3.1; i.e. two years to determine the equilibrium T_g which was then used in a 5-year climatic normal run. Mean annual ground temperature profiles and temperature envelopes predicted for the sites are plotted in Appendix C along with the observed temperatures.

The effect of possible errors in the determination of thermal conductivities was also examined. In Chapter 4, a difference of about 50% between measured and calculated values was observed. The thermal conductivity of the surface layer at two sites, NW-5 with 15 cm of peat and NW-1 with 3 m of sand, was arbitrarily changed $\pm 50\%$ and simulations with climatic normal data performed. The conductivity of the second soil layer (clay) at NW-1 was increased by 50% in a separate run. Results of these simulations are plotted in Figure 5.6. (No data are shown for NW-5, discussed in section 5.5, as very little change occurred).

5.4.2 Modelling Natural Climatic Conditions over the Observation Period

The 5-year climatic normal runs were followed by a simulation of the two years of weather conditions recorded over the observation period at Norman Wells. These 2 years were in fact combined with the climatic normals into one computing run so as to avoid some of equilibrium problems discussed in 5.3.1. The object of this simulation was to determine if the changes in mean temperatures and envelopes observed from 72/73 to 73/74 could be attributed to natural variations in climate. Results of runs for NW-7 and FGH-3 are plotted in Figure 5.7.

Figure 5.6 a) Simulations of ground temperatures at NW-1, 5 yr climatic normal runs for various thermal conductivities of the surface sand layer

Figure 5.6 b) Simulations of ground temperatures at NW-1, 5 yr climatic normal runs for various thermal conductivities of the underlying clay

Figure 5.7 Simulations of climatic conditions of 1972-1973 and 1973-1974 at a) FGH-3 b) NW-7

5.4.3 Microclimatic Effects

Ideally this phase of the analysis would involve the assignment to each site of its respective surface characteristics for 5-year climatic normal simulation runs. This would be followed by simulation runs incorporating the natural climatic variations over the observation period and any change to surface conditions resulting from the measurement program. However, the very general site descriptions available, lacking in detail as to the nature of surface vegetation, surface moisture and snowcover conditions, and their seasonal variation, made it impossible to adequately characterize individual sites. Furthermore, many sites were covered by a dense forest prior to drilling; forest microclimatology is complex, and energy exchanges on the forest floor are difficult to model. In addition the influence of nearby roads, clearings or seismic trails on wind, radiation and snow accumulation for example, are not readily taken into account in the one-dimensional model used here.

Parameter sensitivity analyses (Riseborough, 1982; Burn et al., 1982) have indicated that changes to snowcover thickness and surface soil moisture have the most pronounced effect on the surface energy balance and hence on the surface temperature. The effect of variations in these two factors, such as might reasonably occur in the study area, was thus examined. Site NW-1 was selected for the analysis using the run established for the lithology study. Site wetness was varied over the range of 0.4 to 0.8, and snowfall over the range of 0.5 to 1.5, and the changes in surface temperature determined. Table 5.3 lists the changes in surface conditions examined and the corresponding variation in T_s . Mean annual ground temperatures and envelopes for several of these runs are plotted in Figure 5.8.

TABLE 5.3

EFFECT OF CHANGES OF SNOWFALL AND WETNESS
ON SURFACE TEMPERATURES AT NW-1

RUN	T _s
normal [*]	-2.53
wetness = .8	-4.19
wetness = .7	-3.51
wetness = .4	-2.11
snowfall x .5	-4.16
snowfall x .75	-3.29
snowfall x 1.25	-1.92
snowfall x 1.5	-1.22
snowfall x 1.5 and wetness = .7	-2.16

* Wetness for this run was 0.5 and snowfalls were climatic normals reduced by 33%; snowfalls in subsequent runs are multiples of those in the 'normal' run.

Figure 5.8 Mean ground temperature profiles and envelopes simulated at NW-1, by varying surface wetness and snowfall

5.5 DISCUSSION

The results of each modelling stage were compared to the observed ground temperatures in terms of range of mean annual surface or near-surface temperatures, active layer depth, mean ground temperatures at depth (10 m) and ground temperature envelopes. In this way some gauging of how much of the observed variations in ground temperatures can be accounted for by 1)lithology, 2)natural climate variations and 3)microclimate conditions (including perhaps site disturbances) could be made.

Table 5.4 summarizes the most important information from the analysis. Intuitively, active layer thickness under given climate conditions should be most sensitive to lithologic variations (i.e. to changes in physical and thermal properties of the substrate). Indeed, the range of active layer depths predicted by the lithologic modelling accounts for a large part of that observed. The microclimate variation modelling resulted in little change in active layer depth, as did the modelling of the climate over the observation period.

Surface and near-surface temperatures should be most sensitive to variations in site microclimates. As depth below surface increases, ground temperatures are influenced by an increasingly large area of surface conditions and temperatures become less and less sensitive to localized changes in surface conditions and more sensitive to lithology differences. Recent changes in climate (e.g. on the order of a few years), on the other hand, would be reflected in near-surface temperature variations.

Table 5.4 can now be re-examined within this perspective.

TABLE 5.4

COMPARISON OF OBSERVED AND SIMULATED CONDITIONS

RANGE FOR ALL SITES

	Lithology Modelling	Microclimate Modelling	Observations
Active layer (m)	0.95 to 1.65	1.1 to 1.3	0.5 to 1.6
Surface temp.(°C)	-3.21 to -1.87	-4.16 to -1.22	
0.8 m temp.(°C)	-3.09 to -1.57	-3.75 to -1.27	-2.3 to 0.8
10 m temp.(°C)	-2.23 to -1.36	-3.10 to -0.80	-2.3 to -0.9

5.5.1 Near-Surface Temperatures

The differences in lithology at the sites does not account entirely for the range of mean near-surface temperatures observed. Variations in snowcover and wetness result in a wider range of mean near-surface and surface temperatures, yet still do not account for all of the observed variation.

Figure 5.9, which compares a plot of the range in observed mean ground temperature profiles at the NW and FGH sites with a plot of the range of the mean profiles predicted by lithology modelling, suggests that the predicted surface temperatures cannot fully account for the observed values because the sites have undergone warming at the surface and the mean annual profiles are disturbed. Modelling of the natural climate variations over the observation period indicates that this is not the source of the warming. The simulations of the climate conditions in 72/73 produce a near-surface warming of mean ground temperatures which is smaller in amplitude and shallower in depth of penetration than that observed. The predictions for 73/74 show a pronounced mean cooling (down to a maximum depth of 5 m) and a reduction in minimum temperatures. The influence of this colder year is evident on the measured profiles (see Appendix A2). Cooling is not observed at FGH-2, where the record stopped after one year, nor at NW-2, probably due to a lack of sensors.

It is possible that the warming trend may be the result of increases in air temperatures prior to 1972. By arbitrarily starting the analysis with climatic normal profiles, fluctuations prior to 1972 are ignored. Figure 3.7 indicated that the running mean air temperature (using mean monthly values over 12 month intervals) had increased prior to the observation period. Figure 5.10 shows a more detailed plot of the running mean temperatures

Figure 5.9 Range of temperatures covered by mean annual temperature profiles, both measured and simulated by lithology modelling a) Fort Good Hope sites b) Norman Wells sites

Figure 5.10 Norman Wells air temperatures 1967-1974:
mean annual, mean winter, mean summer
and running mean

starting in 1967. Mean values for the winter months (Nov-Apr), for the summer months (May-Oct), and for each calendar year are also plotted. No linear trend in winter or summer temperatures is evident. Running means vary cyclically, and an increase since late 1971 (i.e. mean of period from early 1971 to early 1972) is noted. Prior to the observation period the temperature was not on average much warmer than the 30-year climatic normal used in MICRO. It is therefore likely that the observed surface temperature disturbance resulted from a disturbance to the local surface conditions.

A possible source of these local microclimatic changes could be the drilling operation, which often involved, for example, the removal of some vegetation cover. Little or no evidence of warming was observed in the mean profiles at NW-5 and NW-7; possibly these sites did not undergo any radical surface disturbance due to drilling.

Simple calculations were performed, based on the method outlined by Carslaw and Jaeger (1959) for determining the temperature changes at depth resulting from a step change in surface temperature, to determine how long ago the surface disturbance might have occurred. The calculations take into account neither the annual wave nor latent heat and work with a mean profile. These calculations showed, for example at NW-1, that a step increase in T_s of 2 K could within a month produce the disturbances observed in the mean profile.¹ The propagation of the temperature change is probably too

1. The calculations were also made for FGH-1 where the observed mean temperatures were disturbed to greater depths - a step change in T_s of 2 K could have occurred here 2 years before drilling. The case of FGH-1 is unique as the site is in addition most likely influenced by noticeable local disturbances - thawing and erosion of the river bank and gullies. Possibly, the forest fire in 1969 had a warming effect at this site.

rapid as the increase most likely was not a step function. However, the order of magnitude is probably correct since very little latent heat effects need be incorporated at the site modelled. Thus site disturbances resulting from drilling could possibly have had a net effect of increasing the ground temperatures.

5.5.2 Deep Temperatures

Lithology modelling has shown that the range of 10 m mean temperatures observed cannot be accounted for solely by subsurface lithology. Possible 50 % errors in thermal conductivity values would have little effect on this predicted range. Changes in conductivity of the surface peat layer at NW-5 had virtually no effect on ground temperatures, presumably because the layer is very thin and its original conductivity very low. The changes to the conductivity of a surface sand layer at NW-1 resulted in less than 0.25 K change in mean temperatures (Figure 5.6a) with corresponding changes to the width of the temperature envelope; e.g. narrower the lower the conductivity and the closer to 0°C. Changes to the conductivity of the thicker second clay layer at NW-1 had a more pronounced effect on the ground temperature regime (Figure 5.6b).

In general, changes in conductivity affect most minimum temperatures which in turn, give rise to changes in the mean. Maximum temperatures change very little because of latent heat effects; the initial profile with $T_s = -2.5$ and $g = 60 \text{ mK m}^{-1}$, is such that most soil water is in the solid state. Thus decreases in mean temperature are not retarded by phase change effects whereas increases are. The position of the maximum of the temperature envelope in the upper 5 m is closer to the freezing point and changes in diffusivity resulting from conductivity changes are overshadowed by the

apparent heat capacity term resulting from latent heat absorption or release. Thus the temperature maxima do not change substantially in the simulations.

Slight changes in snowcover characteristics and/or site wetness could quite easily account for the range in observed 10 m mean ground temperatures. In addition, the individual differences between the lithology modelling temperature predictions and observed temperatures at each site (both in terms of mean temperatures and ground temperature envelopes) could also be accounted for. The microclimate modelling results plotted in Figure 5.8 show the effect of the proximity of the ground temperature to 0°C on the size of the temperature envelope. "The width of the maximum-minimum temperature envelope is narrower the closer the mean annual temperature is to the freezing point" (Goodrich, 1982), because latent heat effects dominate for a proportionately greater fraction of the annual cycle. In addition the depth of zero annual amplitude decreases as the envelope width decreases.

In Figure 5.9a the Fort Good Hope measurements are noticeably warmer than the predictions. These warmer observations may not be due entirely to differences in surface characteristics. In discussing the nature of the ground thermal regime in Chapter 2, the moderating effect of nearby bodies of water was mentioned. Table 5.5 gives a general idea of the proximity of each site to bodies of water and of the percentage of the surrounding area covered by water bodies. The colder profiles predicted by MICRO (Section 5.4.1) at FGH-1, FGH-2 and FGH-4 could be partly explained by the fact that the simulations did not take into account the effect of surrounding lake-covered terrain. Similarly, at NW-7 the influence of the large lake to the south might account for the warmer observed ground temperatures. At NW-6, which is within a low lying swamp, the MICRO predictions would result in even colder

TABLE 5.5

WATER BODIES IN SURROUNDING AREAS

Site	% of surrounding area		Distance to nearest large body (m)
	0.5 km radius	1 km radius	
FGH-1	25	25	50-100
FGH-2	10	12	50
FGH-3	0	5	2000
FGH-4	40	30	100
NW-1	5-10	5-10	1000
NW-2	0	5	900
NW-3	0	5	1500
NW-4	35	25	30
NW-5	0	2	1000
NW-6	0	0	swamp
NW-7	40	25	70

ground temperatures if the correct surface wetness were assigned. In low lying areas with standing water there may be a significant amount of heat contributed via ground water flow (an effect which would not be taken into consideration in MICRO) and perhaps greater accumulation of snow, both of which could elevate the mean annual temperature.

The simulations at NW-1, showing the sensitivity of the energy exchange at the site to changes in wetness and snowcover, clearly indicate that for accurate microclimatic and ground thermal regimes simulations a thorough knowledge of the surface conditions at a site are imperative.

5.6 SUMMARY

The computer simulations of microclimatic and ground thermal regimes described in this chapter indicate that the variations in subsurface lithologies at the study sites could account for a large part of the differences in ground temperatures observed. The differences can also be readily explained by possible site variations in snowcover and/or wetness. The simulations of natural climate conditions over the observation period do not fully account for the observed disturbances to mean annual profiles, nor explain why their "effect" varied, e.g. why some sites showed very little or no warming in 72/73. Surface temperature warming observed at most sites is probably due to man-induced disturbances resulting from the drilling program. Unfortunately data on site specific characteristics are lacking prior to drilling and only sporadic after, and the exact nature of the changes in surface conditions is not known. A reconstruction of the surface conditions and climate, both during the observation program, and for several years prior to the drilling would be necessary to properly resolve the effects of possible site disturbances on the ground thermal regime.

Computer simulations should be utilized with care at all times in order to avoid unrealistic and unstable predictions. As the quote by Nakano and Brown at the beginning of chapter 5 suggests, the modelling has stressed the relative importance of certain physical parameters, and indicated that in future studies detailed measurements of surface, as well as subsurface, conditions should be made prior to and during observation programs. Adequate background data are needed to provide the framework for understanding ground thermal regimes. These measurements should include careful descriptions of the vegetation layer, soil moisture variations throughout the year, snowcover characteristics and local climate conditions (or a knowledge of how these might differ from a nearby AES weather station, e.g. wind, radiation and precipitation regimes).

Where mean annual ground temperature profile predictions are of concern, 50% errors in soil thermal conductivity can be tolerated; if ground temperature envelope predictions are important, then these errors may be too large. Accuracy in estimates of surface parameters is more critical. Even if the thermal properties of the subsurface are accurately measured, unless the site specific characteristics are known, the prediction of T_g will not be accurate and consequently, the prediction of the temperature envelopes at depth may be a poor estimation of the ground temperatures prevailing in the study region. The active layer depth is not as sensitive to these inaccuracies, but the attenuation and depth of penetration of the annual surface temperature wave are more sensitive because of apparent heat capacities and apparent diffusivities arising from latent heat effects. Unless the model can reproduce the measured ground thermal regime while providing an understanding of the controlling parameters, predictions of how a site might behave in the future will not be accurate.

CHAPTER 6

SUMMARY AND CONCLUSIONS

6.1 SUMMARY

Shallow ground temperature data from 11 sites in the Norman Wells and Fort Good Hope region were collected from 1971 to 1974 during research into the environmental effects of the proposed Mackenzie Valley gas pipeline. The study area is located in the discontinuous permafrost zone, where mean annual surface temperatures are generally negative, permafrost thicknesses vary from less than 50 to 80 m, and fine-grained soils are sensitive to slight changes in ground temperatures. A knowledge of the response of ground thermal regimes to natural or man-induced surface changes has important implications for northern development. The present research undertook to analyze the ground thermal regime over the observation period at the study sites, which represent a variety of surface and subsurface conditions. The thesis research could perhaps be described as a "retrofitting" exercise; it dealt with a limited body of available data and attempted to extract information in order to resolve the effects of subsurface lithology, natural climate conditions, microclimatic setting and surface disturbances on the observed ground thermal regime at each site.

The analysis required a knowledge of the annual ground temperature variations, of the thermal properties of the earth materials over a range of positive and negative temperatures, and of the many factors influencing the surface energy balance and its seasonal variation.

The first component of the research (Chapter 3) consisted of synthesizing the ground temperature data to characterize the annual temperature variations at each site. A quality control of the temperature logs was performed, and gaps in the temperature record filled in through the use of curve fitting routines. Temperatures were then interpolated at regular intervals, to enable calculation of mean ground temperatures and ground temperature envelopes.

A diversity of ground temperature conditions was apparent. Mean annual temperature profiles, generated from the interpolations at the study sites, nearly all show signs of a recent increase in surface temperature (of up to 2.5 K). Mean temperatures below the active layer increased at all but three sites from the first to the second year of observation; mean surface temperatures increased at some, decreased at others. In general, ground temperatures were warmer at the FGH sites.

Soil physical and thermal properties were examined in Chapter 4. Published stratigraphic logs and geotechnical data were used to divide each site into representative lithological layers and to calculate for each layer, porosity, volumetric water content, degree of saturation, and percentages of mineral and organic matter. A few measurements of thermal conductivity at negative temperatures were available; data on soil thermal properties was otherwise lacking.

Calculations of thermal diffusivity based on the series of ground temperature observations and on the soil physical properties were thus attempted. Numerical techniques involving the ground temperature data and either 1) solving the Fourier transient heat conduction equation (4.7) for diffusivity or 2) analyzing the phase lag and amplitude attenuation of the annual surface temperature wave (equations 4.9 and 4.10) proved unsuccessful

in systematically generating diffusivities for the soil and temperature conditions at the study sites.

Thermal diffusivities at several temperatures were therefore calculated from the ratio of thermal conductivity to heat capacity; these properties were in turn determined from the soil physical properties. Conductivity calculations made using Johansen's formula were 60 % higher on average than measured conductivities, and calculated thermal conductivities were decreased by 50 % for use in the numerical simulations.

In order to try to understand the reason for the diversity of ground thermal conditions observed, the interactions between climate, microclimate, lithology and ground thermal regime at each site were examined via a computer simulation model (Chapter 5). Application of the one-dimensional computer model required modifications both to the computer program and to the modus operandi, in order to ensure adequate reproduction of snowcover conditions, to avoid disequilibrium temperature profile predictions and to avoid spurious large temperature inversions. The effect of lithology was modelled by imposing the same surface conditions at each site and climatic normals for 5 year simulations. The natural climatic variations of the observation period were then imposed on the above stabilized profiles. Finally the ground thermal response to variations in surface wetness and snowcover was determined for climatic normal conditions.

6.2 CONCLUSIONS

1. Discontinuous ground temperature records may be treated with Fourier series expansion curve fitting routines to continuously generate data and to facilitate the determination of mean ground temperature profiles and temperature envelopes.

2. Temperature data - from fine-grained soils with a high degree of saturation and subjected to an annual freeze-thaw cycle - cannot be systematically used in numerical techniques to calculate soil thermal diffusivity. These techniques are hampered by 1) the irregular spacing of temperature measurements both in time and with depth, 2) soil inhomogeneity and the lack of more than one temperature sensor within each soil layer, 3) heat transfer processes other than pure conduction - such as heat transfer due to moisture flow, and 4) latent heat effects which give rise to temperature-dependant diffusivities.

3. In the absence of laboratory measurements, thermal properties of such soils must be determined from calculations using the soil physical properties. These calculations require a knowledge, or estimates, of volume fraction of soil constituents, grain size characteristics, soil freezing characteristic curves and mineralogy (in order to estimate soil solids thermal conductivity, K_s). The suite of measured physical properties should then include: gravimetric water content, dry bulk density, particle density, volumetric water content (ideally a soil freezing characteristic curve), grain size distribution, mineralogy and organic matter content. These determinations should be made for each soil layer.

4. An average difference of 60% between calculated and measured thermal conductivity may have resulted from errors in selection of K_s , errors in calculation of degree of saturation, errors in selection of soil freezing characteristic curves, or some combination of the above. 50 % errors in thermal conductivity calculations may be tolerated in numerical simulations of mean ground temperature profiles; however, ground temperature envelope predictions may be poor, particularly for soils where latent heat effects predominate in the annual cycle.

5. Numerical modelling revealed that variations in site lithology combined with variations in site wetness and /or snowcover characteristics could account for the entire range of observed ground temperatures. The very general site descriptions made it impossible to adequately characterize individual site microclimates, and thus to quantify these differences. Warmer temperatures at FGH sites may thus be due to a possible thicker snowcover and/or to the influence of surrounding water bodies.

6. Ground temperature warming observed at many sites cannot be explained by the natural climate conditions prior to and during the observation period, and is probably related to man-induced disturbances associated with the drilling program. To properly resolve the effects of possible site disturbances, the numerical modelling should reconstruct the site specific characteristics and climate both before and after drilling. The necessary descriptions of vegetation, and measurements of soil moisture, snowcover conditions and local climate conditions, are however lacking.

7. Monitoring of the ground thermal regime in relation to development projects such as the Norman Wells - Zama pipeline would benefit from inclusion of systems to continuously measure soil moisture profiles (e.g., by Time Domain Reflectometry and neutron probe techniques), snowcover characteristics (e.g., density and thickness throughout a grid), and the surface energy exchange components (e.g., installation of a portable micrometeorology station). The controlling parameters of the ground thermal regime might then be understood, and predictions of future site behaviour in response to the project possible.

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APPENDIX A

GROUND TEMPERATURE DATA

A1

TYPICAL COMPUTER OUTPUT FROM THE

MULTIPLE HARMONIC FOURIER SERIES

EXPANSION PROGRAM: NW-7, 0.8 and 1.5 m

DEPTH = .8 METERS TIME AT 6 1972

NUMBER OF SAMPLES = 45

NUMBER OF TERMS = 20

A(0) = -4.19711

SUM = .26802

DATE	TIME	TEMP(MEAS)	TEMP(CALC)	RESIDUAL
6/ 7/72	.0	-4.40	-4.49	.09
25/ 8/72	50.0	-4.21	-4.19	.02
25/ 9/72	81.0	-4.01	-4.01	.00
24/10/72	110.0	-4.05	-4.14	.09
13/11/72	130.0	-4.11	-4.04	.07
19/11/72	136.0	-4.10	-4.19	.09
26/11/72	143.0	-4.12	-4.22	.10
9/12/72	156.0	-4.16	-4.17	.01
27/12/72	174.0	-4.04	-4.16	.12
7/ 1/73	185.0	-4.90	-4.02	.88
28/ 1/73	206.0	-4.75	-4.41	.34
3/12/72	150.0	-4.19	-4.01	.18
16/12/72	163.0	-4.24	-4.54	.30
3/ 2/73	212.0	-5.01	-4.90	.11
10/ 2/73	219.0	-5.51	-5.62	.11
17/ 2/73	226.0	-6.27	-6.05	.22
25/ 2/73	235.0	-6.12	-6.19	.07
2/ 3/73	239.0	-6.06	-6.16	.10
13/ 3/73	248.0	-6.21	-6.17	.04
28/ 3/73	265.0	-6.23	-6.12	.11
2/ 4/73	270.0	-5.87	-5.93	.06
6/ 4/73	274.0	-5.70	-5.61	.09
13/ 4/73	281.0	-4.09	-5.05	.96
27/ 4/73	295.0	-3.96	-3.73	.23
12/ 5/73	310.0	-1.75	-2.00	.25
24/ 6/73	353.0	-4.01	-4.01	.00
7/ 7/73	366.0	-4.21	-4.25	.04
23/ 7/73	382.0	-6.01	-5.57	.44
6/12/73	518.0	-4.24	-4.10	.14
31/ 1/74	574.0	-5.43	-5.63	.20
2/ 2/74	576.0	-5.60	-5.89	.29
11/ 2/74	585.0	-7.27	-7.01	.26
19/ 2/74	592.0	-7.73	-7.62	.11
25/ 2/74	599.0	-7.86	-7.96	.10
4/ 3/74	606.0	-7.86	-8.14	.28
11/ 3/74	613.0	-8.20	-8.21	.01
19/ 3/74	621.0	-8.25	-8.19	.06
28/ 3/74	630.0	-7.68	-7.94	.26
1/ 4/74	634.0	-7.54	-7.75	.21
9/ 4/74	642.0	-7.27	-7.14	.13
15/ 4/74	643.0	-6.61	-6.44	.17
22/ 4/74	652.0	-5.70	-5.31	.39
9/ 5/74	672.0	-5.10	-5.33	.23
6/ 6/74	700.0	-4.94	-4.09	.85
6/ 7/74	730.0	-4.58	-4.67	.09

* MINIMUM TEMP MEASURED : -8.50
 * MAXIMUM TEMP MEASURED : .65

SINE COEFF.	COS COEFF.	C(J)	FHI(J)
.10183	-.42661	.58609	2.38603
1.08650	1.42450	3.39937	1.13607
-.18134	.16671	.24785	5.82627
-1.17453	1.01466	1.62908	5.38475
-.13543	-.00925	.13274	4.64419
-.02524	-.43656	.43729	3.19934
-.10309	-.17136	.19947	3.67991
-.07720	-.06986	.10412	3.97685
.09103	.07563	.11835	.87256
.20455	-.05117	.21085	1.81594
-.02610	-.06982	.07454	3.49929
.01443	.05501	.05687	.25658
.02311	.10020	.10783	.22664
-.00298	.07864	.07869	6.24524
-.03533	-.05284	.06356	3.73888
-.05812	-.05373	.08052	3.94800
-.02059	.07808	.03481	5.65046
.00844	.02918	.03037	.28163
-.03971	-.05001	.06386	2.47055
-.03300	-.06756	.07519	3.59593

PREDICTED TEMPERATURES, DEPTH - .8 METERS
TIME = 0 AT 6/ 7/1972

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DATE	TIME (DAYS)	TEMP
13/ 7/72	7.0	1.14
20/ 7/72	14.0	1.30
27/ 7/72	21.0	1.47
3/ 8/72	28.0	1.63
10/ 8/72	35.0	1.80
17/ 8/72	42.0	1.96
24/ 8/72	49.0	2.13
31/ 8/72	56.0	2.29
7/ 9/72	63.0	2.46
14/ 9/72	70.0	2.62
21/ 9/72	77.0	2.79
28/ 9/72	84.0	2.95
5/10/72	91.0	3.12
12/10/72	98.0	3.28
19/10/72	105.0	3.45
26/10/72	112.0	3.61
2/11/72	119.0	3.78
9/11/72	126.0	3.94
16/11/72	133.0	4.11
23/11/72	140.0	4.27
30/11/72	147.0	4.44
7/12/72	154.0	4.60
14/12/72	161.0	4.77
21/12/72	168.0	4.93
28/12/72	175.0	5.10
4/ 1/73	182.0	5.26
11/ 1/73	189.0	5.43
18/ 1/73	196.0	5.59
25/ 1/73	203.0	5.76
1/ 2/73	210.0	5.92
8/ 2/73	217.0	6.09
15/ 2/73	224.0	6.25
22/ 2/73	231.0	6.42
1/ 3/73	238.0	6.58
8/ 3/73	245.0	6.75
15/ 3/73	252.0	6.91
22/ 3/73	259.0	7.08
29/ 3/73	266.0	7.24
5/ 4/73	273.0	7.41
12/ 4/73	280.0	7.57
19/ 4/73	287.0	7.74
26/ 4/73	294.0	7.90
3/ 5/73	301.0	8.07
10/ 5/73	308.0	8.23
17/ 5/73	315.0	8.40
24/ 5/73	322.0	8.56
31/ 5/73	329.0	8.73
7/ 6/73	336.0	8.89
14/ 6/73	343.0	9.06
21/ 6/73	350.0	9.22
28/ 6/73	357.0	9.39

* MEAN ANNUAL TEMPERATURE : 7.97
* MINIMUM PREDICTED TEMP : 5.24

* MAXIMUM PREDICTED TEMP :		10.61
5/ 7/73	364.0	10.27
12/ 7/73	371.0	10.44
19/ 7/73	378.0	10.60
26/ 7/73	385.0	10.77
2/ 8/73	392.0	10.93
9/ 8/73	399.0	11.10
16/ 8/73	406.0	11.26
23/ 8/73	413.0	11.43
30/ 8/73	420.0	11.59
6/ 9/73	427.0	11.76
13/ 9/73	434.0	11.92
20/ 9/73	441.0	12.09
27/ 9/73	448.0	12.25
4/10/73	455.0	12.42
11/10/73	462.0	12.58
18/10/73	469.0	12.75
25/10/73	476.0	12.91
1/11/73	483.0	13.08
8/11/73	490.0	13.24
15/11/73	497.0	13.41
22/11/73	504.0	13.57
29/11/73	511.0	13.74
6/12/73	518.0	13.90
13/12/73	525.0	14.07
20/12/73	532.0	14.23
27/12/73	539.0	14.40
3/ 1/74	546.0	14.56
10/ 1/74	553.0	14.73
17/ 1/74	560.0	14.89
24/ 1/74	567.0	15.06
31/ 1/74	574.0	15.22
7/ 2/74	581.0	15.39
14/ 2/74	588.0	15.55
21/ 2/74	595.0	15.72
28/ 2/74	602.0	15.88
7/ 3/74	609.0	16.05
14/ 3/74	616.0	16.21
21/ 3/74	623.0	16.38
28/ 3/74	630.0	16.54
4/ 4/74	637.0	16.71
11/ 4/74	644.0	16.87
18/ 4/74	651.0	17.04
25/ 4/74	658.0	17.20
2/ 5/74	665.0	17.37
9/ 5/74	672.0	17.53
16/ 5/74	679.0	17.70
23/ 5/74	686.0	17.86
30/ 5/74	693.0	18.03
6/ 6/74	700.0	18.19
13/ 6/74	707.0	18.36
20/ 6/74	714.0	18.52
27/ 6/74	721.0	18.69

* MEAN ANNUAL TEMPERATURE : 13.13
* MINIMUM PREDICTED TEMP : 8.22
* MAXIMUM PREDICTED TEMP : 18.64
4/ 7/74 728.0 18.89

DEPTH = .76 M TIME = 0 AT 6/ 7/1972
 TIME INCREMENTS ALONG THE LENGTH AT 8 DAYS PER LINE

TEMPERATURE(C)

INTERVAL = 7. DAYS TOTAL # OF PREDICTIONS = 105 # OF PREDICTIONS IN A YEAR = 52 # OF RUNNING MEANS = 53

RUNNING MEAN TEMPERATURE(D)

DEPTH = 1.5 METERS TIME=0 AT 6/ 7/1972

NUMBER OF SAMPLES = 45

NUMBER OF TERMS = 20

S/C = -4.09777

SUM = .09339

DATE	TIME	TEMP(MEAS)	TEMP(CALC)	RESIDUAL
6/ 7/72	.0	-.96	-.99	.03
25/ 8/72	50.0	-.73	-.73	.00
25/ 8/72	81.0	-.53	-.55	.02
24/10/72	110.0	-.64	-.63	-.01
13/11/72	130.0	-.44	-.44	.00
18/11/72	136.0	-.44	-.41	-.03
26/11/72	143.0	-.44	-.44	.00
9/12/72	156.0	-.42	-.43	.01
27/12/72	174.0	-.41	-.53	.12
7/ 1/73	185.0	-1.33	-1.27	-.06
28/ 1/73	206.0	-2.89	-2.82	-.07
14/12/72	163.0	-.42	-.37	-.05
3/ 2/73	212.0	-3.21	-3.17	-.04
10/ 2/73	219.0	-3.58	-3.63	.05
17/ 2/73	226.0	-4.07	-4.07	.00
26/ 2/73	235.0	-4.48	-4.43	-.05
2/ 3/73	239.0	-4.50	-4.51	.01
11/ 3/73	248.0	-4.65	-4.62	-.03
28/ 3/73	265.0	-4.89	-4.93	.04
2/ 4/73	270.0	-4.90	-4.89	-.09
4/ 4/73	274.0	-4.76	-4.78	.02
17/ 4/73	281.0	-4.51	-4.46	-.05
27/ 4/73	298.0	-3.80	-3.92	.12
12/ 5/73	310.0	-3.62	-3.38	-.24
24/ 5/73	324.0	-2.00	-2.22	.22
24/ 6/73	353.0	-1.22	-1.29	.07
7/ 7/73	366.0	-1.98	-.95	-.03
23/ 7/73	382.0	-.78	-.81	.03
6/12/73	518.0	-.33	-.40	.07
31/ 1/74	574.0	-2.35	-2.58	.23
2/ 2/74	576.0	-2.64	-2.78	.14
11/ 2/74	585.0	-3.98	-3.79	-.19
18/ 2/74	592.0	-4.58	-4.55	-.03
25/ 2/74	599.0	-5.02	-5.08	.06
4/ 3/74	606.0	-5.30	-5.36	.06
11/ 3/74	613.0	-5.57	-5.52	-.05
19/ 3/74	621.0	-5.83	-5.72	-.11
28/ 3/74	630.0	-5.84	-5.92	.08
1/ 4/74	634.0	-5.80	-5.92	.12
9/ 4/74	642.0	-5.77	-5.71	-.06
15/ 4/74	648.0	-5.57	-5.39	-.18
22/ 4/74	655.0	-4.88	-4.98	.10
9/ 5/74	672.0	-4.15	-4.10	-.05
6/ 6/74	700.0	-2.04	-2.15	.11
6/ 7/74	730.0	-1.61	-1.64	.03
* MINIMUM TEMP MEASURED :		-5.84		
* MAXIMUM TEMP MEASURED :		-.33		

SINE COEFF.	COS COEFF.	C(J)	PHI(J)
-.05616	-.27164	.27065	3.38976
2.14000	.18914	2.14834	1.48264
.07143	.06010	.09335	.87132
-.44484	.88534	.99081	5.81758
-.06824	.06745	.09595	5.49197
-.13530	-.15405	.20503	3.86727
-.04377	-.09162	.10153	3.58725
-.05109	.02445	.05684	5.15876
.03240	.05873	.06708	.50414
-.02739	-.02111	.03458	4.05573
-.03672	-.03388	.04996	3.96713
.00851	-.03711	.03807	2.91606
.03625	.03336	.04926	.82686
.00057	-.01074	.01076	3.08891
-.05059	-.01916	.05410	4.35037
.01359	-.04311	.04584	2.79462
.00920	.00128	.00929	1.43246
.05809	.02960	.06520	1.09952
-.00438	-.02245	.02287	3.33424
-.01632	.03012	.03426	5.78651

PREDICTED TEMPERATURES, DEPTH = 1.5 METERS
TIME = 0 AT 6/ 7/1972

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DATE	TIME(DAYS)	TEMP °
6/ 7/72	0	-1.00
13/ 7/72	7.0	-.94
20/ 7/72	14.0	-.89
27/ 7/72	21.0	-.83
3/ 8/72	28.0	-.83
10/ 8/72	35.0	-.80
17/ 8/72	42.0	-.77
24/ 8/72	49.0	-.74
31/ 8/72	56.0	-.69
7/ 9/72	63.0	-.64
14/ 9/72	70.0	-.59
21/ 9/72	77.0	-.54
28/ 9/72	84.0	-.53
5/10/72	91.0	-.54
12/10/72	98.0	-.59
19/10/72	105.0	-.62
26/10/72	112.0	-.62
2/11/72	119.0	-.57
9/11/72	126.0	-.48
16/11/72	133.0	-.42
23/11/72	140.0	-.42
30/11/72	147.0	-.46
7/12/72	154.0	-.45
14/12/72	161.0	-.39
21/12/72	168.0	-.38
28/12/72	175.0	-.57
4/ 1/73	182.0	-1.02
11/ 1/73	189.0	-1.61
18/ 1/73	196.0	-2.18
25/ 1/73	203.0	-2.64
1/ 2/73	210.0	-3.05
8/ 2/73	217.0	-3.50
15/ 2/73	224.0	-3.95
22/ 2/73	231.0	-4.31
1/ 3/73	238.0	-4.50
8/ 3/73	245.0	-4.58
15/ 3/73	252.0	-4.69
22/ 3/73	259.0	-4.85
29/ 3/73	266.0	-4.93
5/ 4/73	273.0	-4.81
12/ 4/73	280.0	-4.51
19/ 4/73	287.0	-4.18
26/ 4/73	294.0	-3.95
3/ 5/73	301.0	-3.78
10/ 5/73	308.0	-3.50
17/ 5/73	315.0	-3.01
24/ 5/73	322.0	-2.39
31/ 5/73	329.0	-1.87
7/ 6/73	336.0	-1.57
14/ 6/73	343.0	-1.46
21/ 6/73	350.0	-1.36
28/ 6/73	357.0	-1.19
* MEAN ANNUAL TEMPERATURE		-1.92 *
* MINIMUM PREDICTED TEMP :		-4.93

* MAXIMUM PREDICTED TEMP :			-.38
5/ 7/73	364.0		-1.00
12/ 7/73	371.0		-.87
19/ 7/73	378.0		-.82
26/ 7/73	385.0		-.81
2/ 8/73	392.0		-.77
9/ 8/73	399.0		-.71
16/ 8/73	406.0		-.67
23/ 8/73	413.0		-.67
30/ 8/73	420.0		-.68
6/ 9/73	427.0		-.65
13/ 9/73	434.0		-.60
20/ 9/73	441.0		-.56
27/ 9/73	448.0		-.56
4/10/73	455.0		-.55
11/10/73	462.0		-.53
18/10/73	469.0		-.48
25/10/73	476.0		-.45
1/11/73	483.0		-.44
8/11/73	490.0		-.44
15/11/73	497.0		-.41
22/11/73	504.0		-.36
29/11/73	511.0		-.34
6/12/73	518.0		-.40
13/12/73	525.0		-.56
20/12/73	532.0		-.81
27/12/73	539.0		-1.11
3/ 1/74	546.0		-1.37
10/ 1/74	553.0		-1.59
17/ 1/74	560.0		-1.78
24/ 1/74	567.0		-2.07
31/ 1/74	574.0		-2.58
7/ 2/74	581.0		-3.32
14/ 2/74	588.0		-4.14
21/ 2/74	595.0		-4.01
28/ 2/74	602.0		-5.23
7/ 3/74	609.0		-5.43
14/ 3/74	616.0		-5.59
21/ 3/74	623.0		-5.77
28/ 3/74	630.0		-5.92
4/ 4/74	637.0		-5.88
11/ 4/74	644.0		-5.62
18/ 4/74	651.0		-5.22
25/ 4/74	658.0		-4.83
2/ 5/74	665.0		-4.47
9/ 5/74	672.0		-4.10
16/ 5/74	679.0		-3.63
23/ 5/74	686.0		-3.07
30/ 5/74	693.0		-2.54
6/ 6/74	700.0		-2.15
13/ 6/74	707.0		-1.92
20/ 6/74	714.0		-1.81
27/ 6/74	721.0		-1.73
* MEAN ANNUAL TEMPERATURE			-2.17 *
* MINIMUM PREDICTED TEMP :			5.92
* MAXIMUM PREDICTED TEMP :			-1.34
4/ 7/74	728.0		-1.66

DEPTH = 1.50 M TIME = 0 AT 6/ 7/1972
 TIME INCREMENTS ALONG THE LENGTH AT 8 DAYS PER LINE

INTERVAL = 7. DAYS TOTAL # OF PREDICTIONS = 105 # OF PREDICTIONS IN A YEAR = 52 # OF RUNNING MEANS = 53

RUNNING MEAN TEMPERATURE(C)

MEAN -2.18

-2.03

1.89

INTERVAL

A2

OBSERVED MEAN ANNUAL GROUND TEMPERATURES
AND TEMPERATURE ENVELOPES OBTAINED FROM
THE FOURIER SERIES EXPANSIONS AT EACH SITE

FORT GOOD HOPE SITES

— 12/71 to 11/72
 --- 7/72 to 6/73
 7/73 to 6/74

A3

GEO THERM PLOTS

A4

RUNNING MEAN TEMPERATURES

(52 week periods)

7/73-6/74

APPENDIX B

THERMAL PROPERTIES

B1

DIFFUSIVITIES CALCULATED FROM THE ATTENUATION
OF THE AMPLITUDE OF THE ANNUAL TEMPERATURE WAVE

Legend: Year 1 refers to annual wave of 72/73

Year 2 refers to annual wave of 73/74

"Measured" refers to calculations using the observations

"Interpolated" refers to calculations using the data
generated from the Fourier series expansion

$$\% \Delta = \frac{\text{Measured} - \text{interpolated}}{\text{Measured}} \times 100$$

TABLE B1 DIFFUSIVITIES CALCULATED FROM THE ATTENUATION
OF THE AMPLITUDE OF THE ANNUAL TEMPERATURE WAVE

SITE	LAYER	SENSOR PAIR	YEAR	DIFFUSIVITY ($10^{-6} \text{m}^2 \text{s}^{-1}$)		% Δ	TEMP RANGE
				MEASURED	INTERPOLATED		
FGH-2	2	0.8-1.5	1	0.51	0.56	10	-6.2, 0.0
		2.3-3.0	1	1.20	1.08	-10	-2.9, -0.8
	3	3.0-4.6	1	0.70	1.13	60	-2.8, -1.0
		4.6-6.1	1	0.45	0.54	20	-2.4, -1.3
		2.3-6.1	1	0.65	0.81	25	-2.9, -0.8
	4	7.6-9.1	1	0.91	0.67	-26	-1.7, -1.4
FGH-3	2	3.0-4.6	1	1.17	1.55	32	-2.0, -0.5
			2	1.29	7.68	500+	-2.4, -0.4
		4.6-6.1	1	0.58	0.97	67	-1.8, -0.9
			2	1.13	2.71	140	-2.1, -0.7
		3.0-6.1	1	0.86	1.22	42	-2.0, -0.5
			2	1.29	4.34	236	-2.4, -0.4
	3	7.6-9.1	1	0.14	0.28	100	-1.9, -1.4
			2	0.26	0.82	215	-1.7, -1.3
FGH-4	1	0.8-1.5	1	0.09	0.10	10	-3.2, 1.6
			2	0.10	0.10	0	-4.8, 1.8
	2	3.0-4.6	1	0.13	0.09	-27	-1.4, -0.8
			2	0.13	0.17	30	-1.7, -0.7
NW-1	1	0.8-1.5	1	0.07	0.07	0	-7.9, 6.2
			2	0.06	0.07	17	-10.7, 6.3
		1.5-2.3	1	1.08	1.05	-3	-5.6, 0.0
			2	0.40	0.40	0	-6.8, 0.0
		0.8-2.3	1	0.17	0.14	-18	-7.9, 6.2
			2	0.13	0.18	38	-10.7, 6.3
	2	4.6-6.1	1	0.28	0.27	-4	-3.2, -1.5
		6.1-7.6	1	0.49	1.27	159	-2.6, -1.5
		7.6-9.1	1	0.21	0.32	52	-2.3, -1.9
		4.6-9.1	1	0.29	0.43	48	-3.2, -1.5
NW-2	2	1.5-2.3	1	1.46	1.62	10	-3.4, -0.9
			2	1.54	1.64	7	-3.0, -0.6
		2.3-3.0	1	0.66	0.52	20	-3.4, -1.4
			2	0.81	0.76	-6	-3.0, -1.1
		3.0-4.6	1	0.48	0.59	-18	-3.2, -1.7
			2	0.51	0.43	-15	-2.9, -1.4
		4.6-6.1	1	0.42	0.43	2	-2.9, -2.2
			2	0.37	0.28	-24	-2.6, -1.9
		6.1-7.6	1	0.43	0.33	-23	-2.7, -2.3
			2	0.29	0.29	0	-2.4, -2.2
		1.5-7.6	1	0.33	0.45	36	-3.4, -0.9

TABLE B1 (cont'd)

SITE	LAYER	SENSOR PAIR	YEAR	DIFFUSIVITY ($10^{-6} \text{m}^2 \text{s}^{-1}$)		% Δ	TEMP RANGE
				MEASURED	INTERPOLATED		
NW-3	1	0.8-1.5	1	0.21	0.19	-10	-5.1, 0.5
			2	0.10	0.09	-10	-7.7, 1.7
		1.5-3.0	1	0.27	0.25	-7	-3.8, -0.5
			2	0.42	0.40	-5	-4.8, -0.3
		3.0-4.6	1	0.60	0.57	-5	-2.4, -1.1
			2	0.53	0.65	23	-3.1, -0.9
		4.6-6.1	1	0.22	0.29	32	-2.1, -1.6
		0.8-4.6	2	0.30	0.33	10	-5.1, -1.6
		0.8-6.1	1	0.30	0.23	-23	-5.1, 0.5
NW-4	2	0.8-1.5	1	0.12	0.09	-25	-4.7, 0.6
			2	0.06	0.03	-50	-5.5, 1.9
		1.5-2.3	1	0.36	1.16	222	-3.1, -0.8
			2	0.19	0.32	68	-2.4, -0.4
		2.3-3.0	1	0.29	0.14	-52	-2.8, -0.9
			2	0.65	0.43	-34	-2.1, -0.9
		0.8-2.3	1	0.19	0.22	16	-4.7, 0.6
			2	0.10	0.08	-20	-5.5, 1.9
	3	3.0-4.6	1	0.39	0.33	-15	-2.6, -1.6
NW-5	2	0.8-1.5	1	0.10	0.10	0	-6.5, 3.0
			2	0.08	0.08	0	-8.6, 3.3
		1.5-2.3	1	0.19	0.20	5	-4.8, -0.3
			2	0.22	0.25	14	-5.4, -0.3
		0.8-2.3	1	0.13	0.15	15	-6.5, 3.0
			2	0.13	0.14	8	-8.6, 3.3
	3	4.6-6.1	1	0.70	0.72	3	-3.2, -1.7
			2	0.68	0.62	9	-3.4, -1.7
	4	7.6-9.1	1	0.12	0.60	400	-2.4, -1.9
			2	0.12	0.34	183	-2.5, -1.9
NW-6	1	0.8-1.5	1	0.02	0.02	0	-1.7, 4.6
			2	0.05	0.06	-17	-7.6, 5.0
		1.5-2.3	1	0.16	0.19	-19	-1.3, -0.3
			2	0.28	0.30	7	-4.9, -0.4
		0.8-2.3	1	0.04	0.04	0	-1.7, 4.6
			2	0.10	0.11	-9	-7.6, 5.0
NW-7	2	1.5-2.3	1	0.65	1.22	88	-4.9, -0.4
			2	0.69	0.69	0	-5.9, -0.3
		2.3-3.0	2	0.73	0.60	-18	-4.8, -0.6
		1.5-3.0	2	0.71	0.65	-8	-5.8, -0.3
	3	6.1-7.6	1	0.32	0.28	-13	-1.7, -1.0
			2	0.28	0.22	-21	-1.9, -1.1
		7.6-9.1	1	0.30	0.19	-37	-1.3, -1.0

B2

CALCULATED HEAT CAPACITIES
AND THERMAL CONDUCTIVITIES

TABLE B2 HEAT CAPACITY (C), THERMAL CONDUCTIVITY (K)
AND THERMAL DIFFUSIVITY (κ)

<u>SITE</u> <u>LAYER</u>	<u>MATERIAL</u>	<u>T</u> (°C)	<u>θ</u> (%)	<u>Sr</u>	<u>C</u> (MJm ⁻³ K ⁻¹)	<u>K (Wm⁻¹K⁻¹)</u> Johan Meas.		<u>κ</u> (10 ⁻⁶ m ² s ⁻¹)
FGH-1 1	peat	0	43	0.5	2.18	0.37		0.08
		-1	35		2.00			0.10
		-2	15		1.35	0.45		0.17
		-10	10		1.23	0.49		0.20
FGH-1 2	sand	0	39	1.0	2.85	2.42		0.42
		-0.5	13		2.26	3.44		0.76
		-1	11		2.12	3.53		0.80
		-2	9		2.17	3.63		0.82
FGH-1 3	silt	0	41	1.0	3.29	1.85	1.02	0.31
		-0.5	16		2.33	2.60	1.44	0.62
		-1	12		2.24	2.74	1.52	0.68
		-2	10		2.19	2.81	1.53	0.71
FGH-1 4	sand	0	47	2.0	3.02	2.00	1.60	0.53
		-0.5	10		2.18	3.30	2.38	1.09
		-1	8		2.14	3.39	2.45	1.00
		-2	7		2.11	3.44	2.46	1.17
FGH-1 5	silt	0	34	0.8	2.54	1.56	1.12	0.44
		-0.5	16		2.13	1.97	1.72	0.81
		-1	12		2.04	2.08	1.82	0.89
		-2	10		2.00	2.13	1.87	0.94
FGH-2 1	peat	0	43	0.5	2.18	0.37		0.08
		-1	35		2.00	0.38		0.10
		-2	15		1.35	0.45		0.17
		-10	10		1.23	0.49		0.20
FGH-2 2	silty clay	0	62	1.0	3.46	1.19		0.17
		-0.5	30		2.72	1.84		0.34
		-1	20		2.49	2.11		0.42
		-2	17		2.42	2.19		0.45
FGH-2 3	till	0	17	0.7	2.12	2.04	0.89	0.40
		-0.5	16		2.19	1.89	0.97	0.44
		-1	12		2.10	2.00	1.02	0.49
		-2	10		2.05	2.05	1.05	0.51
FGH-2 4	shale	0	18	1.0	2.26	1.26	0.97	0.43
		-1	12		2.13	1.33	1.25	0.59
		-2	10		2.08	1.37	1.27	0.61
FGH-3 1	silty clay	0	50	1.0	3.14	1.51		0.24
		-0.5	30		2.68	1.98		0.37
		-1	20		2.45	2.26		0.46
FGH-3 2	gravel	-2	10	1.0	2.38	2.36		0.50
		0	37		2.80	2.57	1.46	0.52
		-1	5		2.08	3.92	2.23	1.07
		-2	3		2.03	4.04	2.40	1.18

TABLE B2 (cont'd)

<u>SITE LAYER</u>	<u>MATERIAL</u>	<u>T</u> (°C)	<u>θ</u> (%)	<u>Sr</u>	<u>C</u> (MJm ⁻³ K ⁻¹)	<u>K</u> (Wm ⁻¹ K ⁻¹)		<u>κ</u> (10 ⁻⁶ m ² s ⁻¹)
						Johan	Meas.	
FGH-3 3	clay	0	35	1.0	2.77	1.96	0.97	0.35
		-0.5	30		2.66	2.09	1.04	0.39
		-1	20		2.43	2.40	1.19	0.49
		-2	17		2.36	2.49	1.24	0.53
FGH-4 1	silty sand	0	41	1.0	2.90	2.05		0.35
		-0.5	13		2.26	3.00		0.66
		-1	11		2.12	3.08		0.70
		-2	9		2.17	3.17		0.73
FGH-4 2	clay- ice	0	85	7.1	3.87	0.75	0.6	0.16
		-0.5	8		2.10	2.16		0.51
		-1	5		2.03	2.23		0.55
		-2	4		2.02	2.24	1.18	0.58
NW-1 1	sand	0	39	1.0	2.86	2.42		0.42
		-1	5		2.08	3.80		0.91
		-2	3		2.03	3.90		0.96
NW-1 2	clay	0	49	1.0	3.28	1.54	0.99	0.30
		-0.5	30		2.64	1.99	1.28	0.48
		-1	20		2.41	2.28	1.46	0.61
		-2	17		2.34	2.37	1.53	0.65
NW-2 1	silt- org.	0	94	7.2	4.07	0.64	0.6	0.15
		-1	7		2.05	2.08		0.51
		-2	3		1.98	2.22		0.56
		-10	2		1.95	2.23	1.5	0.77
NW-2 2	clay	0	33	0.5	2.11	0.81		0.16
		-0.5	30		2.23	0.94		0.21
		-1	20		1.81	1.07		0.30
		-2	17		1.74	1.12	1.46	0.84
NW-3 1	till	0	45	0.6	3.25	1.05		0.16
		-0.5	16		2.08	1.45		0.35
		-1	12		1.99	1.53		0.38
		-2	10		1.95	1.57		0.40
NW-3 2	shale	0	18	1.0	2.26	1.23	0.50	0.22
		-1	12		2.13	1.33		0.27
		-2	10		2.08	1.37	0.60	0.29
NW-4 1	peat	0	95	1.4	4.12	0.55		0.07
		-1	30		2.62	1.33		0.25
		-2	10		2.16	1.70		0.39
		-10	7		2.09	1.80		0.43
NW-4 2	clay	0	56	0.9	3.47	1.18		0.17
		-0.5	30		2.41	1.68		0.35
		-1	20		2.18	1.93		0.44
		-2	17		2.12	2.01		0.47
NW-4 3	clay	0	41	0.8	3.21	1.40		0.22
		-0.5	30		2.37	1.65		0.35
		-1	20		2.14	1.87		0.44
		-2	17		2.07	2.01		0.50

TABLE B2 (cont'd)

SITE LAYER	MATERIAL	T	θ	Sr	C (MJm ⁻³ K ⁻¹)	K (Wm ⁻¹ K ⁻¹)		κ (10 ⁻⁶ m ² s ⁻¹)
		(°C)	(%)			Johan	Meas.	
NW-5	peat	0	43	0.5	2.18	0.37		0.08
1		-1	35		2.00	0.38		0.10
		-2	15		1.35	0.45		0.17
		-10	10		1.23	0.49		0.20
NW-5	till	0	40	1.0	2.88	1.84		0.32
2		-0.5	35		2.77	1.97		0.36
		-1	20		2.42	2.41		0.50
		-2	17		2.35	2.51		0.53
NW-5	shale	0	24	1.0	2.41	1.16		0.24
3	weath.	-0.5	13		2.15	1.35		0.32
		-1	11		2.11	1.39		0.33
		-2	9		2.06	1.43		0.35
NW-5	shale	0	19	1.0	2.29	1.22		0.27
4		-0.5	13		2.15	1.32		0.31
		-1	11		2.12	1.36		0.32
		-2	9		2.06	1.40		0.34
NW-6	peat	0	85	1.0	3.03	0.50		0.06
1		-1	35		2.80	1.01		0.18
		-2	15		2.34	1.33		0.28
		-10	10		2.22	1.42		0.32
NW-6	clayey	0	53	1.0	3.19	1.44		0.51
2	silt	-0.5	16		2.34	2.38		0.51
		-1	12		2.25	2.51		0.56
		-2	10		2.20	2.58		0.59
NW-6	siltstone	0	22	1.0	2.36	1.19		0.25
3		-1	11		2.11	1.38		0.33
		-2	9		2.06	1.41		0.34
NW-6	shale	0	19	1.0	2.29	1.22		0.27
4		-1	12		2.13	1.34		0.31
		-2	10		2.08	1.46		0.35
NW-7	silt-	0	62	1.0	3.39	1.21		0.18
1	org	-1	35		2.77	1.75		0.31
		-2	15		2.31	2.30		0.50
NW-7	sand	0	19	0.7	2.22	2.32		0.52
2		-0.5	13		2.08	2.19		0.53
		-1	11		2.05	2.25		0.56
		-2	9		1.99	2.31		0.58
NW-7	shale	0	19	1.0	2.29	1.22		0.27
3		-1	12		2.13	1.34		0.31
		-2	10		2.08	1.46		0.37

Heat capacity calculated from physical properties; thermal conductivity,
Johan = calculated with Johansen's formula, Meas = laboratory measurement.

APPENDIX C

GROUND TEMPERATURES PREDICTED BY
LITHOLOGY SIMULATIONS
(Mean ground temperature and envelope
plotted with observed temperatures)

(m)

(m)

— SIMULATION
-- 72/73 ... 73/74 (C)

(m)

END

0	6	H	0	9	H	8	4
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FIN

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