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Practical limits to mapping the active layer:

A geostatistical evaluation of frost table probe and remote sensing data

by

David Viljoen, B.Sc.

Department of Geography

Carleton University

Ottawa, Canada

December 13, 1993

A thesis submitted to the faculty of Graduate Studies in partial fulfillment of the requirements for the degree of Master of Arts, Geography.

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A geostatistical evaluation of frost table probe and remote sensing data."

submitted by David Warren Viljoen, B.Sc. (Geography), in partial fulfillment of the requirements for the degree of Master of Arts (Geography).

A handwritten signature in dark ink, appearing to read "Michael W. Smith", written over a horizontal line.

Thesis Supervisor

Michael W. Smith

A handwritten signature in dark ink, appearing to read "Iain Wallace", written over a horizontal line.

Chairman, Geography Department

Iain Wallace

Carleton University,

January 6, 1994

Abstract

Detailed active layer measurements were made on Richards Island, Northwest Territories with two main objectives:

- 1) to understand relations between the active layer depth and surface temperature, surface cover, macro- and micro-topography, and snow depth;
- 2) to quantify the spatial variability of the active layer and related phenomena in order to assess the utility of various measuring approaches including frost table probe, ground probing radar, aerial photography, and satellite imagery.

The depth of active layer exhibits very high spatial variability due to thermal properties of the ground materials, microtopography, and surface cover. According to the results, accurate mapping of the active layer in areas of continuous permafrost requires sampling intervals that make it impractical to map large areas. Geostatistics, and semivariograms in particular, are valuable for estimating the spatial variability of natural phenomena, and may be used to design efficient and effective sampling schemes.

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Our kind and generous neighbours John and Simone Carr for the use of their printer in times of desperate need.

My wife Anne for her support and love. This thesis was my endeavour but it is our accomplishment. It simply would not have been achieved without her. Thanks to my children Joshua and Mikaela for their spirited support. To them the word "thesis" conjures up images of something that makes Daddy hide in his room for extended periods of time.

After recently watching Walt Disney's Alladin, Joshua said "If I had a Genie, my first wish would be that your thesis would be all done." Thankfully, that wish has now come true.

David Viljoen, December 1993.

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1. Introduction

1.1. An introduction to permafrost

Permafrost refers to ground that remains below a temperature of 0°C for more than one full year (Williams, 1982). The term permafrost implies that the ground has been, and will be, permanently frozen. This is not the case. Any factor, above or below the surface, that affects the ground thermal regime can influence permafrost conditions.

The linkage between climate and ground temperatures has been well documented (Brown, 1978; Harris, 1981; Williams & Smith, 1989). The relationship is complicated by several factors such as thermal properties of the ground, surface cover, snow cover, hydrology, and topography. Where these factors combine to produce a mean annual ground temperature below 0°C, permafrost will exist. Over time scales of decades or millennia, consistent exposure to sub-zero temperatures can lead to substantial aggradation of permafrost. In the high Arctic, borehole data have shown permafrost thicknesses measured in hundreds of metres (Judge, 1986). Alternatively, climatic change, migration of surface water bodies, or alteration of surficial cover can lead to degradation of permafrost. The thermal conductivity, heat capacity, thermal diffusivity, and thermal admittance of the surficial materials govern the sub-surface thermal regime and transfer of heat (Oke, 1987). These will ultimately determine the impact of changes at the surface on the underlying permafrost.

1.2. Importance of permafrost and global change

General Circulation Models (GCMs) are complex models of the climate system based on the fundamental physical laws that govern the Earth's climate. Through simplification and parameterization of factors that influence climate, simulations of the features of global warming are attempted, for example. The magnitude of predicted changes in climate for a

given variation of a parameter, such as increased concentration of carbon dioxide in the atmosphere, will vary from model to model. This is primarily due to the differences in the representation of effects at the wide range of spatial and temporal scales that must be considered (Gates, 1981; Stuart & Judge, 1991). However, most models consistently show that global warming will be two to three times greater in the Arctic than in the more temperate regions south of 60° N degrees latitude (Figure 1.1). It seems useful, therefore, to evaluate the potential of monitoring and modeling phenomena in the Arctic regions for assessing climate change as well as providing more accurate input to global climate models. This is important from both the scientific and practical point of view.

From the scientific point of view there is still much that needs to be learned with respect to the role of Arctic regions in global climate. In response to this, the CRYSYS (CRYospheric SYStem) Project was created as part of the National Aeronautics and Space Agency's (NASA) Earth Observing System (EOS) in support of the Global Change Mission to the Planet Earth Program. The main goal of CRYSYS is to develop an information system for understanding the sensitivity of high latitude ecology, cryosphere, and physiography to climate change. It will "rely mainly on satellite observations of northern regions as well as ground climate and cryospheric observations" (Simard, 1992).

From the practical point of view, climate change in northern environments will have dramatic effects. Areas underlain by ice-rich permafrost may experience substantial subsidence should the ice melt, while in coastal areas loss of soil mechanical strength due to thawing will cause relatively rapid shoreline regression. This could affect existing man-made structures and will have to be considered for future developments such as pipelines, roads, harbours or new buildings.

The problem becomes one of identifying phenomena that could be used to monitor the progress of global warming as well as providing input to other monitoring and modeling

programs. Permafrost is widespread at high latitudes and has several qualifiable and quantifiable properties that could potentially be used for these purposes.

Figure 1.1 Estimates of past and future changes of surface temperature in response to global warming (Kellogg, 1979).

1.3. Qualifiable and quantifiable properties of permafrost

Frozen ground has several properties that may be observed and/or measured, and which may prove useful in assessing the impact of climate change on soil temperature, periglacial phenomena, and the active layer.

In terms of temperature monitoring, one of the most useful properties of frozen ground compared to unfrozen ground is the relative stability of the ground's thermal properties below the frost table. The primary reason for this is a difference in moisture regime. In frozen soils, moisture and phase changes are not a factor below the frost table. In unfrozen soil, moisture, phase changes, and water migration have a dramatic effect on the ground thermal regime in the upper 20 metres.

In a frozen soil with excess ice, the thermal conditions are dominated by the thermal properties of the ice. These soils are sensitive to temperature changes since ice has a relatively low heat capacity and high conductivity. The combination of stable thermal properties below the frost table and temperature sensitivity are ideal properties for measuring small changes in temperature that can be expected below the depth penetration of annual temperature fluctuations. As ice approaches the freezing point, however, temperature fluctuations are attenuated by the diversion of energy to the latent heat of melting. For this reason, soil that is well below freezing is better for monitoring temperature changes.

Periglacial features and processes can be used as indicators of the presence or absence of frozen ground. Examples of periglacial features are patterned ground, pingos, ice wedges, and thermokarst. Periglacial processes include thaw landslides, solifluction, thaw settlement, and frost heaving. In response to climatic warming, periglacial features such as pingos or ice wedges can be expected to experience some degree of melting with associated changes in morphology. The problem with using such features is that any such change will have a significant time lag due to the enormous latent heat of fusion that would have to be overcome for morphological changes to occur. Changes in patterned ground and thermokarst development are similar to ice wedges and pingos in that there is a significant time lag in response to climatic change. In addition, remnants of such features can persist long after the ground has thawed which often makes it difficult to determine remotely which features are active and which are relict. Periglacial processes could be used as indicators of climate change by relating the rates of such processes to climate. One of the problems with this approach is that many variables including material, water content, and topography can also affect the rates of these processes which are themselves difficult to measure quantitatively in the field.

It has been suggested by many observers that the annual depth of thaw (active layer) could be used as a means of monitoring climatic change. One of the advantages of using the depth of active layer is that it has distinct properties that make it easy to measure in the field. Mechanically, the difference in strength between frozen and unfrozen materials is dramatic and can be easily assessed using a frost table probe. Electrically and electromagnetically, the difference in the conductive and dielectric properties of unfrozen and frozen soil is significant and can be measured as a strong reflection by ground probing radar. Active-layer development is related to properties that can be assessed remotely such as vegetative cover, moisture conditions, and topography. Relative to the other permafrost characteristics and processes, active-layer depth seems to have more potential in terms of mapping the response of permafrost to climate change.

1.4. Context of this work

Many attempts at measuring and mapping permafrost and its characteristics have been attempted ranging from manual probing (Mackay, 1977) and ground probing radar (Annan, 1975; Dallimore, 1992) to satellite imagery (Bélanger, 1990; Schreier, 1981; Morrissey, 1986). Considerable effort has also been expended on examining and modeling relationships of permafrost to climate (Harris, 1981; Williams, 1982), microclimate (Smith, 1975), and a variety of terrain characteristics (Peddle, 1992; Jorgenson & Kreig, 1988; Brown, 1978). These studies have spanned a wide range of scales from site specific to regional, but none attempt to quantify or estimate the spatial variability of permafrost conditions.

Figure 1.2 illustrates the potential pitfalls of various sampling and interpolation methods. Figure 1.2 a) shows a resampled version of the "real" data. Here the pixels shown are much larger than the measured real phenomenon for display purposes. For the plan views shown in Figure 1.2 b) and c), a small sampling of the real data was taken and used to

generate Thiessen polygons and an isoplethic map based on traditional contouring techniques. Transects across each of the resulting plan views were taken and plotted. Comparing the transects from a), b), and c) shows clearly the problems associated with sparse sampling schemes and inappropriate interpolation.

Figure 1.2 Plan views and transects (profiles) highlighting the problems associated with inadequate sampling for mapping natural phenomena. (a) "Real data; (b) Thiessen Polygons; (c) contours from linear interpolation.

Maps are abstractions of reality and must rely on a certain amount of smoothing and generalization. However, many geomorphological phenomena exhibit high spatial variability that is unmappable at common scales (e.g. 1:50,000). The challenge is to determine the scale of variability to assess the inapplicability of naive averaging, trend surface fitting, or any other interpolation technique (Culling, 1986).

Geostatistical methods have been used for more than 20 years to quantify the spatial variability of natural phenomena (Matheron, 1971; Journel & Huijbregts, 1978; McBratney & Webster, 1981; Curran, 1988). However, an on-line computer search of more than 500,000 bibliographic entries in a geoscientific database failed to reveal one reference to the application of geostatistics to the spatial variability of permafrost or related phenomena (refer to Appendix 7.1 for details of the on-line search).

1.5. Objectives

Mapping of permafrost conditions could be useful in providing:

- a means to monitor climatic change and,
- insight into the relative stability of surficial materials.

In terms of climatic change, the monitoring and modeling of permafrost conditions could be used as a barometer of annual and multi-year variations in climatic conditions. The stability of surficial materials is related to such factors as the susceptibility to slumping (thaw settlement), sliding (thaw landslides), and solifluction (slow downslope movement).

The problem of mapping permafrost conditions is primarily one of scale. This involves providing answers to such questions as:

- What permafrost condition or parameter varies at useful mappable scales (e.g. 1:50,000 or smaller)?
- Can this parameter be used to provide a map of:
 - a) areas which might be useful in monitoring climatic change ?

b) the relative stability of surficial materials ?

It has already been suggested that the active layer might provide the answers to such questions. However, the spatial variability of the active layer must be assessed as it relates to the resolution of various measuring tools including remote sensing devices.

In this regard, data were collected on Richards Island, Northwest Territories, with the following objectives:

- to assess the variability of the active layer and its relationship to other variables such as surface temperature, surface cover, proximity effects, and ground materials based on manual probing of the active layer
- to quantify the spatial variability of the active layer and assess the utility of data provided by remote measuring devices such as ground probing radar (GPR), aerial photography, and satellite imagery
- to assess the practical limits to mapping the active layer by quantifying its spatial variability as it relates to various measuring devices including frost table probes, ground probing radar, aerial photography, and satellite imagery.

1.6. Overview

Chapter 2 will provide background on the development of the active layer as it relates to the surface energy balance and ground thermal properties. The theory and capabilities of the various measurement devices, and the concepts behind geostatistical analysis are also discussed. Chapter 3 will present the methods used in the collection and processing of the data and Chapter 4 the results of the analysis and a discussion of their relevance. Chapter 5 will summarize the fundamental conclusions of this work and suggest future research possibilities.

2. Background

The active layer is a spatially heterogeneous phenomenon that depends primarily on the surface energy balance and the thermal properties of the ground. Many types of remotely sensed and derived data may be used to infer surface cover and subsurface materials. This chapter will discuss the active layer, its relationship to the surface energy balance, and the time scales involved in the response of the ground thermal regime to changes in the surface energy balance. Components of the surface energy balance will be examined in some detail to highlight the important factors relevant to the ground thermal regime. Properties of the ground will be discussed to illustrate how the surface energy balance is manifest in the subsurface environment. Background will be provided on measuring instruments in the context of providing information on parameters relevant to the surface energy balance or ground thermal properties. These instruments include frost-table probes and tubes, ground probing radar, scanned aerial photography, and the Thematic Mapper sensor on the Landsat series of Earth-observation satellites. The theory and utility of semivariograms will be discussed in relation to quantifying the spatial variability of natural phenomena and to the data provided by these instruments that attempt to measure them.

2.1. Thermal conditions in permafrost

Figure 2.1 shows a schematic of an idealized subsurface temperature profile. If the climatic conditions do not exhibit any trend on time scales measured in 10's or 100's of years towards increasing or decreasing temperature, then the temperature at the surface can be thought of as a long term steady-state condition with a certain amount of year-to-year variation (ΔT_s). Associated with each annual average temperature (T_s) are the seasonal changes that create a temperature amplitude "envelope" that decays exponentially with depth into the ground as described by the following equation:

$$S_z = S_a e^{-z\sqrt{\pi/iat}} \quad (2.1)$$

where S_a = amplitude temperature envelope at the surface for one year (ΔT_a)

S_z = amplitude of S_a at depth z

a = diffusivity of material

t = period of temperature fluctuation (1 year in this case)

Figure 2.1 Schematic of geothermal gradient, permafrost, and active layer.

The depth to which this annual variation is felt can be determined from the following equation:

$$Z_a = 2\sqrt{\pi at} \quad (2.2)$$

The depth at which the envelope maximum crosses the 0°C isotherm defines the active layer depth (Z_t). In a simple world, for each variation in surface temperature from year-to-year there would be an associated change in the depth of the active layer (ΔZ_t). In the

real world, however, a change in annual surface temperature may not produce the expected change in the active layer due to changing thermal properties of the ground. Moisture content is the most important factor in this regard for two reasons. 1) Moisture alters the thermal properties of the active layer by increasing the heat capacity and thermal conductivity. If the underlying permafrost is ice-rich, then the latent heat of fusion will further increase the energy required to deepen the active layer. 2) Water movement in the soil acts as a heat transfer mechanism.

Solving for t in Equation 2.2 provides some idea of the time scales involved in changing thermal conditions at depth. Figure 2.1 shows that a temperature change at the surface will not be manifest at 150 m depth for approximately 114 years depending on the thermal diffusivity of the material (this computation was made using a diffusivity typical of a sandy clay i.e. $5 \times 10^{-7} \text{m}^2/\text{sec}$). Figure 2.2 shows the response of the geothermal gradient to a step change in surface temperature. The thick, straight line represents the initial steady state equilibrium gradient. Note that even after 1000 years, the geothermal gradient has still not reached a steady state equilibrium condition.

If there was a step change in temperature at the surface, it would take thousands of years for the entire geothermal gradient to approach a steady-state equilibrium. Since the time scales involved for modification of the geothermal gradient are so large, paleoclimatic reconstruction is possible by examining such parameters as the depth of permafrost and changes in the slope of the geothermal gradient with depth (Allen, 1988; Judge, 1986). Figure 2.3 shows the thickness of ice bearing permafrost in the Arctic Coastal Plain with an inset map showing the extent of two of the three glaciations that can be detected in this area. There seems to be a correlation between the limits of glaciation and thinner permafrost. Theoretically, the thick glacial ice insulates the ground from the extreme cold climate inhibiting frost penetration. Over thousands of years, the juxtaposition of insulated and exposed ground resulted in a permafrost thickness difference of more than

600 m Now, more than 12,000 years after the ice retreated from this area, the influence of the Quaternary glaciations is still apparent.

Figure 2.2 Sequence of geothermal response to a surface step change (Molochuskin, 1973).

Figure 2.3 Contours showing the thickness of ice bearing permafrost. Inset map shows limits of various glaciations (Judge, 1986).

It is obvious that there should be no concern over the disappearance of permafrost greater than 100 metres thick due to climate change, on human time scales. However, global warming could have a significant impact on near-surface permafrost; particularly the depth of the active layer and the potential melting of shallow ground ice within 20 metres of the surface. This is especially true for areas where ice-rich permafrost is presently near the melting point.

In order to more fully understand active-layer development, the relationship between climate and ground thermal conditions will be examined. This entails a review of the components of the surface energy balance and factors that affect the ground thermal regime.

2.2. Surface energy balance

Annual active layer thickness is a reflection of the surface energy balance and the thermal properties of the ground. The main factors which influence the ground thermal regime are summarized in Table 2.1.

Table 2.1: Factors which influence the surface energy balance.

Factor	Important location-specific aspects
Astronomical	latitude, time of day, time of year
Atmospheric/climatic	temperature, wind velocity, cloud cover, rainfall, snowfall
Topographic	slope, aspect, altitude
Surficial properties	albedo, emissivity, roughness, wetness, vegetative cover
Subsurface properties	bulk density, thermal diffusivity, water content

(Cihlar & McQuinllan, 1978)

These factors may be conveniently divided into those which influence the surface energy balance (i.e. astronomical, atmospheric/climatic, topographic, and surficial properties) and those which affect the subsurface properties which are important to the ground thermal regime (i.e. bulk density, thermal diffusivity, water content). Solution of the surface energy balance provides the equilibrium surface temperature which is an important boundary condition for the ground thermal regime. The subsurface properties determine how the equilibrium surface temperature will influence the depth of thaw, among other things. Throughout this discussion, "positive" values refer to energy fluxes incident on the surface and "negative" are related to energy being reflected or emitted away from the surface.

The surface energy balance accounts for all radiative, conductive, and convective energy fluxes occurring at a surface and is mathematically represented as follows:

$$Q^* + Q_H + Q_{LE} + Q_G = 0 \text{ [MJ m}^{-2} \text{ time}^{-1}] \quad (2.3)$$

where Q^* = net radiation

Q_H = sensible heat flux

Q_{LE} = latent heat flux

Q_G = ground heat flux

Note the units are per unit time. This convention is used since it is equally valid to apply Equation 2.3 for describing the surface energy balance in time scales ranging from seconds to millennia. In order to identify the various factors which influence the surface energy balance, each of the components will be examined individually in a little more detail.

2.2.1. Net radiation

Net radiation is defined as the net exchange between all wave outgoing and incoming radiation (Whittow, 1988). It accounts for all wavelengths of radiation incident on a surface as well as all wavelengths of radiation reflected by or emitted from a surface. Since the radiative energy processes are different for short- and long-wave radiation, it is a common practice to consider each of these portions of the electromagnetic spectrum separately, i.e.:

$$Q^* = K^* + L^* \quad (2.4)$$

where K^* = net short wave radiation [$\text{MJ m}^{-2} \text{ time}^{-1}$]

L^* = net long wave radiation [$\text{MJ m}^{-2} \text{ time}^{-1}$]

Short-wave radiation refers to wavelengths from 0.3 to 3.0 μm (i.e. visible through to reflected infrared) and long-wave radiation refers to wavelengths from 3.0 to 20 μm (thermal infrared band). Figure 2.4 shows the various types of radiation found in the electromagnetic spectrum ranging from Gamma and X-ray to visible and microwave. It highlights the 'so-called' short- and long-wave portions of the spectrum as well as the wavelengths to which the various bands of the Landsat Thematic Mapper sensor are sensitive.

Figure 2.4 The electromagnetic spectrum showing main spectral classes of electromagnetic radiation and wavelengths to which the Thematic Mapper Sensor bands are sensitive.

Equation 2.5 is a more detailed version of equation 2.4 and highlights the major factors used in determining the net radiation balance. After a number of manipulations and substitutions as documented in many climatological texts (Sellers, 1965; Oke, 1987), the more detailed version of the net radiation balance is:

$$Q^* = K\downarrow (1 - \alpha) - \{(1 - bn^2) [\epsilon_s \{ \epsilon_{ao} \sigma T_o^4 - \sigma T_s^4 \}]\} \quad (2.5)$$

where $K\downarrow$ = short wave radiation incident on the surface

α = albedo

b = regression coefficient that accounts for cloud type and height

n = fraction of the sky covered with cloud ($0 \leq n \leq 1$)

ϵ_s = emissivity of the surface

ϵ_{ao} = emissivity of a clear sky

σ = Stefan-Boltzmann's constant

T_o = screen-level temperature

T_s = surface temperature

The emissivity of a cloudless sky is one of the most difficult parameters to quantify in the net radiation balance. Many equations have been developed since the early 1900's which utilize atmospheric temperature and/or water vapour pressure in estimating ϵ_{ao} . An equation with emissivities that compare favourably to those found useful in Arctic environments, and which utilizes both screen-level temperature and vapour pressure, has been developed (Idso, 1981):

$$\epsilon_{ao} = 0.7 + 5.95 \times 10^{-5} \cdot e_o \cdot e^{1500/T_o} \quad (2.6)$$

where e_o = screen-level vapour pressure [mb]

T_o = screen-level air temperature [K]

In order to estimate either short- or long-wave radiative terms, it is often customary to evaluate conditions for cloudless conditions and then correct for cloud (Davies et al, 1975). For example, a total of 65 cloudless days occurred at Resolute during the period November 1963 to December 1972 (Suckling & Wolfe, 1979). All cloudless days occurred during the winter months: November to May. This suggests that cloud cover is

perhaps a more important factor in the Arctic as compared to more southerly regions with respect to influencing the net radiation balance.

2.2.2. Sensible heat flux

In thermodynamics, sensible heat refers to heat transferred to a substance that causes an associated temperature change (Rogers & Mayhew, 1980). Apart from the few millimetres above a surface (i.e. the laminar boundary layer), sensible heat is transferred from the surface by a turbulent flux into the atmosphere. Aerodynamic theory has been used to provide an equation which quantifies this flux:

$$Q_H = \frac{K_H \rho c_p (T_s + T_o)}{z_o} \quad (2.7)$$

where K_H = transfer coefficient evaluated from aerodynamic theory using Dyer-Businger diabatic wind profile [$\text{m}^2 \text{s}^{-1}$]

ρ = density of the air [kg m^{-3}]

c_p = specific heat of air at constant pressure [$\text{J kg}^{-1} \text{K}^{-1}$]

T_s = equilibrium surface temperature [K]

T_o = screen-level temperature [K]

z_o = height above surface at which temperature is given, i.e. 1 m

The transfer coefficient takes into account the surface roughness (aerodynamic roughness length), temperature gradient, wind speed gradient, and atmospheric stability as described in the literature (Oke, 1987, p. 380).

2.2.3. Latent heat flux

Latent heat refers to the amount of energy emitted or absorbed when a substance (water) changes its state or phase without any change of temperature within the substance. As with sensible heat, there is an aerodynamic equivalent for evaluating the latent heat flux. Unfortunately it requires information about the vapour density gradient. Another option which has been used effectively is an equilibrium model described by the following equation (Smith, 1976):

$$Q_{LE} = \frac{\mu \cdot S}{S + \gamma} (Q^* - Q_G) \quad (2.8)$$

where μ = the site wetness factor

S = the slope of the saturation vapour pressure-temperature curve

γ = psychrometric constant, i.e.

$\gamma = C_a/L_v$ [Pa K⁻¹] or [kg m⁻³ K⁻¹]

where C_a = volumetric heat capacity of air [J m⁻³ K⁻¹]

L_v = latent heat of vaporization [J m⁻³] or [J kg⁻¹]

This model is very useful in terms of "user input" since it only requires a site wetness factor, air temperature, Q^* , and Q_G . The volumetric heat capacity of air can be found from:

$$C_a = \rho c_p \quad (2.9)$$

where ρ = density of air [kg m⁻³]

c_p = specific heat of air at constant pressure [kJ kg⁻¹ K⁻¹]

The density and specific heat of air are functions of temperature, pressure, and moisture content (Rogers & Mayhew, 1981).

Effective equations for estimating the latent heat of vapourization of water (L_v) have been developed which require only the screen-level temperature. An equation which incorporates a temperature dependency into the value for the latent heat of vaporization of water is given as (Henderson-Sellers, 1984):

$$L_v(T) = 1.91846 \times 10^6 \{T_o/(T_o - 33.91)\}^2 \text{ [J kg}^{-1}\text{]} \quad (2.10)$$

2.2.4. Ground heat flux

Of the three mechanisms of heat transport; conduction, convection, and radiation, conduction is by far the most important in both the solid and liquid (soil moisture) components of the environment (de Vries, 1975). The equation which describes the conduction of heat in the ground is given by:

$$Q_G = -K_G \frac{T_s - T_z}{z} \quad (2.11)$$

where Q_G = ground heat flux [$\text{MJ m}^{-2} \text{ time}^{-1}$]
 K_G = thermal conductivity of the ground [$\text{MJ m}^{-1} \text{ K}^{-1}$]
 T_s = equilibrium surface temperature [K]
 T_z = temperature at depth z [K]
 z = depth at which T_z is measured

Many factors influence the ability of the ground to conduct heat, as well as the change in temperature that such heat flow might create. These factors will be discussed in the next Section.

2.3. Ground thermal regime

In order to understand the nature of the ground thermal regime, some basic thermal principles must be reviewed. There are three fundamental thermal properties which influence the thermal regime of soils: thermal conductivity, heat capacity, and latent heat. The heterogeneous nature of soils makes for a more variable thermal regime relative to that found in bedrock.

Thermal conductivity is defined as the ability to transmit heat energy within a substance from a point of high temperature to a point of low temperature by molecular impact without overall movement of the substance itself (Whittow, 1984). In soils, thermal conduction is achieved by direct solid-to-solid contact between soil grains or via the soil solution which occupies the pore space between soil particles. If air occupies a large proportion of the pore space, then the conductivity of the soil will be reduced. Figure 2.5 shows the dependency of thermal conductivity on soil moisture content for sand, loam, and clay. Organic matter on the surface, in the form of leaf litter, moss, or peat, or beneath the surface has a low thermal conductivity. The presence of peat in a mineral soil reduces its thermal conductivity and diffusivity in both the frozen and unfrozen states (Farouki, 1986).

Figure 2.5 Influence of moisture content on the thermal conductivity of sand, silt, and clay (Farouki, 1986).

Heat capacity refers to the amount of heat that is required to change the temperature of a unit volume or mass of material by one degree (Whittow, 1984). The heat capacity of a soil can be thought of as the sum of the heat capacities of its constituents in proportion to their relative abundance. It is useful to mention the term "apparent heat capacity" which consists "not only of the true heat capacity of the water, eventual ice, and mineral constituents, but also of that component of heat which is liberated in association with the transformation of water into ice" (Williams, 1982).

Thermal diffusivity is a derived quantity which is defined as the ratio of thermal conductivity to the heat capacity of a material. Since it relates the ability of a material to transmit heat to the amount of heat required for a unit change in temperature, it is a measure of the ease with which a unit volume or mass of material will change temperature with a change in its boundary heat condition.

Latent heat is defined as the amount of energy emitted or absorbed when a body changes its state or phase, without any change of temperature within that body (Whittow, 1984). Since water can be present in all phases in soils, it is necessary to discuss both the latent heat of fusion (0.333 MJ kg^{-1}) and the latent heat of vaporization (2.45 MJ kg^{-1}). The

latent heat of fusion is the amount of heat that must be removed (or added) in order to change water to ice (or ice to water). Similarly, the latent heat of vaporization is the amount of heat that must be removed (or added) to change water from vapour to liquid (or liquid to vapour). This means that in certain situations where changes of state and a heat flux to or from a material are involved, latent heat serves to delay a change of temperature until a change of state is effected.

Ice and water have dramatically different thermal properties. The heat capacity of ice is half that of water yet its thermal conductivity is four times greater (Williams, 1982). This suggests that generalized parameters for thermal conductivity and heat capacity must be created for unfrozen soil as well as for frozen soil at the macroscale. Treating the ground thermal regime as a 'two-layered' system, consisting of a frozen and a non-frozen part both with a unique set of thermal properties, has been used successfully in the past (Smith, 1976; Jackson et al, 1975).

The movement of unfrozen water in a soil represents a mass transfer of heat energy and therefore plays an important role in determining freezing and thawing of soils. Water migration in an unsaturated frozen soil has been found to be directly proportional to the temperature gradient, inversely proportional to the square root of time. It decreases with the decrease in temperature in the power law form, and changes with dry density (Xiaozu et al, 1982). The temperature gradient is the driving force for water migration in frozen soils. A large temperature gradient means that there will be a more rapid migration of unfrozen water. However, such a gradient also leads to rapid freezing of the ground and, consequently, results in an rapid lowering of the permeability. As the temperature decreases, the amount of unfrozen water decreases and the permeability of the soil decreases as well due to the occupation of pore space by ice and trapped air. Therefore, the migration of water decreases "in the power law form" with decreasing temperature. The flux of unfrozen water increases with an increase in dry density to a certain maximum

above which it begins to decrease. When the dry density is low, there are many large pores in the soil so that the continuity of the unfrozen water film among soil particles is poor. Increasing the dry density may offer more opportunities for unfrozen water passing through these paths because of the increasing continuity of the unfrozen water films. Eventually, increasing the dry density truncates the paths for water migration leading to a decrease in the rate of migration (Xiao et al, 1985).

Table 2.2 shows the thermal properties of some common rocks and minerals. The thermal diffusivity of quartz is of particular interest since it is far greater than any other rock or mineral. Knowledge of the soil quartz content is important as a basis for a more accurate estimate of the thermal diffusivity of coarse soils. If the quartz content of the soil is relatively low, then a diffusivity value which approaches that of mica and feldspar (i.e. ≈ 1) can be used. For fine-grained soils water content is the dominant factor that must be considered in determining a diffusivity value. (Farouki, 1982).

Table 2.2 Thermal conductivity, heat capacity, and diffusivity values for some common rocks and minerals (from Williams, 1982).

Mineral	Thermal Conductivity k	Heat Capacity c	Thermal Diffusivity a
	[W m ⁻¹ °C ⁻¹]	[10 ⁶ J m ⁻³ °C ⁻¹]	[10 ⁻⁶ m ² s ⁻¹]
Quartz	8.0	2.1	3.81
Mica	3.0	2.4	1.25
Limestone	2.9	2.5	1.16
Orthoclase feldspar	2.0	2.0	1.00
Granite	2.0	2.1	0.95
Shale	1.5	1.6	0.94

2.4. Surface energy balance summary

All of the components of the surface energy balance may be found directly or indirectly from the list of parameters in Table 2.3. Each of these is variable on different spatial scales and this should be taken into account when considering the surface energy balance. The effect of all these factors is reflected in the active layer.

Table 2.3 Summary of variables important to evaluation of the surface energy balance.

Symbol	Description	Potential Sources
α	albedo	field measurement using spectrophotometers; aerial photographs, satellite imagery
b	regression coefficient that accounts for cloud type and height	field measurement; weather satellites and manned weather stations
e_o	screen-level vapour pressure	field measurement; manned and automatic weather stations
ϵ_s	emissivity of the surface ($0 \leq \epsilon \leq 1$)	field measurement or empirical values from field experiments
$K_{\downarrow}$	short wave radiation incident on the surface	field measurement; manned and automatic weather stations
K_G	thermal conductivity of the ground	field measurement; related to density and soil moisture which can be inferred remotely from Synthetic Aperture Radar(SAR) and satellite imagery
K_H	transfer coefficient evaluated from aerodynamic theory using Dyer-Businger diabatic wind profile	field measurement; related to surface roughness which can be inferred from SAR imagery
n	fraction of the sky covered with cloud ($0 \leq n \leq 1$)	field observations; weather satellites and manned weather stations
T_o	screen-level temperature [K]	field measurement; manned and automatic weather stations
T_s	equilibrium surface temperature	field measurement; inferred from satellite or airborne thermal sensors
T_z	temperature at depth z	field measurement; inferred from K_G and T_s

2.5. Frost table probes and tubes

Frost table probes can be used to accurately measure the depth of thaw. If fitted with a thermistor, the temperature gradient in the active layer can be measured and used to estimate thermal properties of the soil and the effect of various terrain factors on the ground thermal regime. They can be used for rapid reconnaissance since many measurements can be made in a relatively short period of time. It is suited to studies where the relationship of the active layer to surface cover, topography, and other parameters is desired. Throughout this document, the term surface cover will be used as a generic reference to the organic or inorganic materials that cover the ground's surface.

Frost tubes are semi-permanent devices that are used for measuring the year-to-year fluctuations in thaw depth as well as seasonal development of the active layer. They can provide perhaps the most accurate measurements of thaw depth when installed properly

2.5.1. Design of frost table probes and tubes

A frost table probe consists of a metal rod, typically from 0.65 to 1.5 cm in diameter and 1 to 1.5 metres in length, with a sharpened tip on one end and a handle on the other. Some are hollow allowing for installation of a thermistor in the tip to measure the subsurface temperature which can be used to confirm contact with the frost table at 0°C. If many soil temperatures are recorded between the surface and the frost table, then the temperature gradient within the active layer can be estimated.

The difference in mechanical strength between frozen and unfrozen soil material is dramatic in soils ranging from sandy to silty. In silty clays or clays, this difference becomes more difficult to detect due to unfrozen pore water slightly below 0°C. For this reason, larger diameter probes provide a better "feel" for the frost table than small diameter probes (Makay, 1977).

A frost tube actually consists of two tubes. An outer tube acts as a sheath for the interior removable tube. The outer tube is inserted into a hole drilled to a diameter somewhat larger than the tube diameter and to a depth perhaps 75 cm deeper than the frost table. The hole is then back-filled with the original soil material. The interior tube is filled with distilled water and allowed to freeze. A plastic bead is placed in the tube and rests on ice in the frost tube. When the ground begins to re-freeze in the Fall, the bead is frozen at the depth of maximum thaw. Special provision has to be made to account for frost heaving effects on the tube to accurately measure the depth of thaw. Frost tubes must be installed before, and extracted after, the maximum depth of thaw is reached to ensure accurate measurement.

2.5.2. Accuracy of frost table probes and tubes

Field experience with a 1.25 cm frost table probe suggests that measured thaw depths are accurate to within 1 cm in sandy and silty soils and perhaps overestimated by 2-3 cm in silty clays and clays. However, this accuracy is inversely proportional to the diameter of the frost table probe. It is possible to push smaller diameter probes (i.e. 0.65 to 0.95 cm) up to 23 cm below the frost table in fine grained materials such as clay or mud (Mackay, 1977). The accuracy in the estimation of the frost table can be achieved by probing the frost table several times within a few centimeters so that the "feel" of the soil material can be improved. This helps to better distinguish the unfrozen and frozen ground. The goal is usually to measure the depth of maximum thaw so the underlying assumption is that the depth of thaw has reached its maximum at time of probing. Optimally, probing should be done when the active layer is fully developed (i.e. late summer). However, as long as the probing is performed sometime late in the summer, then the error in active layer measurement will be minimal since the depth develops with the square-root of time.

The active layer is influenced by factors within an area at the surface and a volume of material in the ground. The area and volume involved are proportional to the active layer depth. Since the ground surface and frost table are both variable in space, a representative active layer value must involve spatial averaging of these two parameters. Therefore, measurements of both the ground surface and frost table must be made with respect to a standard reference to account for the effects of microtopography on active layer estimation.

2.5.3. Spatial resolution of frost table probes and tubes

Spatial resolution of a frost table probe is dependent on the goals of the researcher. The lowest possible resolution would theoretically be perhaps slightly larger than the diameter of the probe. However, it is not usual practice to ensure that the probe is inserted absolutely perpendicular to the frost table surface so that the practical minimum resolution is probably on the order of a few centimetres.

Spatial resolution of a frost tube is on the same order as the probe but practically speaking, it is not usually possible to drill two holes within 15 cm of each other.

2.5.4. Temporal resolution of frost table probes and tubes

The temporal resolution of a measuring device is important when considering variations in a phenomenon through time. In terms of monitoring an area throughout a given season, the frost table probe requires a high degree of spatial accuracy for mapping or marking of sample sites so that they can be revisited to within 5 cm. Frost tubes have the advantage of continually monitoring a given site. However, the number of sample sites is potentially much greater for the frost table probe than for the tube. For monitoring year-after-year, the frost table probe would require marking sample sites so that the markers could survive extreme weather conditions while not affecting the surface energy balance or ground thermal regime. Although frost tube installation has an immediate impact on the surface

energy balance and the ground thermal regime due to disruption of the surface cover and ground materials, the site usually recovers to a sufficient degree.

2.6. Ground probing radar (GPR)

Ground probing radar, like the frost table probes and tubes, is able to estimate the depth of active layer but does so by measuring the two-way travel time of a reflected electromagnetic pulse. One of the advantages of GPR is that it provides more continuous information than frost table probes in a shorter period of time. However, it does require local ground control since the two-way travel time is sensitive to soil moisture and soil texture.

2.6.1. Theory of GPR

This section will present a simplified review of some of the basic theory behind GPR. For a more complete discussion there are many possible references (e.g. Pilon, 1983). GPR refers to the use of radio waves, with center frequencies typically in the range from 1 to 1000 MHz, to map subsurface structure and features (Annan, 1992). The depth of reflecting features can be computed from the two-way travel time of the electromagnetic wave and the velocity through the intervening medium. The velocity can be computed from the following equation.

$$v = \frac{c}{\sqrt{K}} \quad (2.12)$$

where c = speed of light in a vacuum = 3.0×10^8 m/s
 K = dielectric constant of the medium.

Figure 2.6 a) is a schematic illustration of a GPR survey and Figure 2.6 b) is a graphical display of the resulting reflection section.

(a)

(b)

Figure 2.6 (a) Schematic of GPR survey and (b) graphical display of resulting reflection.

The ability to resolve features using GPR depends primarily on the electromagnetic and electrical properties of relative permittivity (dielectric constant) and conductivity respectively. The difference in dielectric constant between target and background will determine the proportion of signal reflected. Figure 2.7 shows the proportion of energy reflected from the interface in a two-layer system. It is based on the following equation for the reflection coefficient:

$$R = \frac{\sqrt{K_1} - \sqrt{K_2}}{\sqrt{K_1} + \sqrt{K_2}} \quad (2.13)$$

where K_1 = dielectric constant of material 1

K_2 = dielectric constant of material 2

Figure 2.7 clearly shows that for a given subsurface layer (K_2) the reflection coefficient is more sensitive to changes in dielectric constant when the dielectric constant of the top layer (K_1) is smaller (e.g. 30 or less). For example, if an area is underlain by a thick layer of saturated clay ($K_2 = 50$), then the reflection coefficient for a top layer of a partially saturated sand ($K_1 = 20$) would be more than twice that if the same clay was covered with water ($K_1 = 80$) even though the difference in dielectric constant between the top and bottom layers is the same. The implication for earth sciences is that GPR is sensitive to the changes in dielectric constant typical in common soil and bedrock materials and is therefore a practical system for exploring subsurface features. Table 2.4 provides typical electromagnetic and electrical properties of some common geological materials.

Water is an important factor with respect to the dielectric properties of a soil as shown in Figure 2.8. In terms of measuring the active layer, this can be either beneficial or detrimental. If the water is only a thin layer perched on the permafrost, then it can increase the reflection coefficient and improve discrimination of the frost table. If the moisture level in the active layer is high, then the resulting attenuation and reflection of the radar signal will make it more difficult to recognize the return from the frost table.

Figure 2.7 Proportion of energy reflected in a two-layer system (Annan, 1992)

Table 2.4 Typical dielectric content, electrical conductivity, velocity, and attenuation observed in common geological materials at a frequency of approximately 100 MHz (Annan, 1992).

Material	Dielectric constant K	Conductivity (mS/m) σ	Velocity (m/ns) v	Attenuation (dB/m) α
Air	1	0	0.3	0
Ice	3 - 4	0.01	0.16	0.01
Dry sand	3 - 5	0.01	0.15	0.01
Granite	4 - 6	0.01 - 1	0.13	0.01 - 1
Limestone	4 - 8	0.5 - 2	0.12	0.4 - 1
Dry Salt	5 - 6	0.01 - 1	0.13	0.01 - 1
Shales	5 - 15	1 - 100	0.09	1 - 100
Silts	5 - 30	1 - 100	0.07	1 - 100
Clays	5 - 40	2 - 1000	0.06	1 - 300
Saturated sand	20 - 30	0.1 - 1	0.06	0.03 - 0.3
Distilled water	80	0.01	0.033	0.002
Fresh water	80	0.5	0.033	0.1
Sea Water	80	3000	0.01	1000

Figure 2.8 Variation of dielectric constant and electromagnetic wave propagation with liquid water content (Annan, 1992).

The conductivity of a material will affect signal attenuation and, therefore, the ability of the transmitted pulse to propagate in the ground. High conductivities mean that the energy from the transmitting antenna will be dissipated in the form of conduction currents generated in the material. The attenuation can be estimated from the following equation:

$$\alpha = 1.69 \frac{\sigma}{\sqrt{K}} \text{ dB/m} \quad (2.14)$$

The attenuation in the signal often renders GPR ineffective in certain environments. For example, GPR has been used to identify ice-rich sediments or massive ice in the upper 50 m of Quaternary sediments (Dallimore, 1992). Sediments, with a dielectric constant ranging from 5 to 30, underlain by ice, with a dielectric constant of 3 to 4, will usually

result in fairly strong reflections. The conductivities are typically less than 100 mS/m for sediments that are partially saturated with fresh water and 0.01 mS/m for ice. These relatively low conductivities mean that the reflections at the sediment-ice interface will not be attenuated as the electromagnetic wave travels from the transmitting antenna to the receiving antenna. If the sediments are partially saturated with sea water, with a conductivity of 3000 mS/m ($\alpha = 1000$ dB/m), compared to fresh water at 0.5 mS/m ($\alpha = 0.1$ dB/m), the transmitted pulse could be attenuated such that the reflected signal would be below the resolving power of the receiver.

To maximize the signal to the receiving antenna, the antenna separation should be such that the refraction focusing peak in the transmitting and receiving antenna patterns are fixed on a common point at the target depth. Antenna separation is dependent on the dielectric properties of the ground, the antenna pattern, and the depth to the target. Figure 2.9 shows how the antenna pattern changes with dielectric constant. The equation that estimates the optimum antenna separation is given by:

$$S = \frac{2D}{\sqrt{K-1}} [\text{m}] \quad (2.15)$$

where D = the target depth [m]

For a target depth of 0.5 m and dielectric constant of 10, the antenna separation would be about 30 cm.

Figure 2.9 Variation in antenna pattern with changes in dielectric constant (Annan, 1992).

2.6.2. Vertical resolution of GPR

The water and ice that are usually associated with the frost table provide a contrast in dielectric constant that results in a strong reflection when using GPR. The time window, W , is the period for which the receiver will record signals returned from the ground after a pulse is transmitted. This determines the maximum depth of signals that will be recorded and can be computed using the following equation.

$$W = 1.3 \frac{2 \times D}{v} [\text{ns}] \quad (2.16)$$

where D = depth [m]
 v = velocity [m/s]

The factor of 1.3 allows for uncertainties in velocity and depth variations (Annan, 1992).

The system sampling interval determines how many data points will be recorded within the time window. With consideration given to center frequency, system bandwidth, and a "safety margin", maximum sampling intervals can be determined from the equation:

$$\tau = \frac{1000}{6f_c} [\text{ns}] \quad (2.17)$$

where f_c = system centre frequency [MHz]

The spatial resolution puts a lower bound on the centre frequency that should be used in a GPR survey. The avoidance of unwanted radar returns from fine-scale features such as small cracks, joints, or organic matter lenses in the ground places an upper bound on the centre frequency. Another factor which further constrains the upper bound is that the target size should occupy a certain proportion of the radar beam footprint in order for the returned beam to be coherent (so-called Fresnel zone size). This factor is not of concern for probing the frost table since relative to the Fresnel zone size, the frost table is infinite. Given these frequency constraints, the following "quick guide" table can be derived for selecting centre frequency.

Table 2.5 Suggested GPR centre frequency for different target depths (Annan, 1992).

Target depth [m]	Centre Frequency, f_c [MHz]
0.5	1000
1.0	500
2.0	200
5.0	100
10	50
30	25
50	10

Since the target depth ranges from 0.2 to 0.8 m, a centre frequency of 1000 MHz would seem to be optimal. Equation 2.17 suggests a maximum sampling rate of 0.17 ns. Using a

dielectric constant of 10, the velocity of an electromagnetic wave would be 0.095 m/ns (9.5 cm/ns) according to Equation 2.12. Measuring two-way travel times at a sampling rate of 0.17 ns means that the wave could be sampled every 0.81 cm [= (0.17 ns / 2)*9.5 cm/ns] and this represents the theoretical lower limit for such a system. However, given variations in moisture content and other factors that affect the dielectric properties of the soil, the practical lower limit would be higher. Time offsets would have to be applied to account for changes in microtopography which would have to be measured using conventional surveying techniques.

2.6.3. Spatial resolution of GPR

Station spacing and the GPR "footprint" determine the spatial resolution of a GPR survey and both are dependent on the centre operating frequency of the system and the dielectric properties of the ground. The spatial resolution of GPR is linked directly to station spacing (Δx shown in Figure 2.6 a). The station spacing should not exceed the Nyquist interval which means less than one quarter of the wavelength in the subsurface material as given in the following equation:

$$\Delta x \leq \frac{c}{4f\sqrt{K}} = \frac{75}{f\sqrt{K}} [\text{m}] \quad (2.18)$$

where f = frequency [MHz]

Given this relation the station spacing would be less than or equal to 2.4 cm for a frequency of 1000 MHz and a dielectric constant of 10. This station spacing can be achieved with appropriate system settings and dragging the GPR system over the terrain in a continuous profiling mode.

Figure 2.10 provides a simplified view of the "footprint" of the transmitted pulse. The major (A) and minor (B) axes can be computed from:

$$A = \frac{\lambda}{4} + \frac{D}{\sqrt{K-1}} [\text{m}] \quad (2.19)$$

$$B = A / 2 [\text{m}] \quad (2.20)$$

For a frequency of 1000 MHz, a dielectric constant of 10, and a target depth of 0.5 m, these dimensions would be 24 and 12 cm respectively. If the target has a roughness with a spatial frequency higher than 24 cm, then the return signal might be attenuated due to diffuse reflection of the transmitted pulse.

Figure 2.10 Simplified GPR footprint concept with the shaded ellipse depicting an illuminated plane at depth.

2.6.4. Temporal resolution of GPR

As with frost table probes, the temporal resolution of GPR surveys is controlled by the researcher and particular objectives of the study. Timing of the survey with respect to the

thaw season must be considered if the objective is to map maximum depth of annual thaw. If the GPR system is dragged on a sled, there is the potential of surveying several hundred metres per day (Pilon, 1979). If the site is to be revisited year-after-year or several times in a thaw season, however, the impact on the site could begin to affect the active layer through surface cover disturbance. The measured active-layer development, therefore, would actually be more of an impact assessment of GPR surveys.

2.7. Scanned aerial photography

Aerial photographs are a popular product used in many applications that involve surficial mapping such as geology, soils, agriculture, and forestry. This is due to their scale and high spatial resolution. In the context of permafrost studies, aerial photographs clearly show features such as patterned ground, pingos, thaw slumping, surface morphology, and surface cover. The challenge is to relate features visible on aerial photographs, particularly surface cover, and radiometric variations within a given feature, to the thaw depth variability.

2.7.1. Theory of aerial photography

Aerial photographs, like all imaging systems that rely on reflected light, depend on a few fundamental phenomena. The Sun, with a surface temperature of 5900°C, emits most of its radiant power in the visible range of the electromagnetic spectrum. This supplies all of the energy required for imaging in the visible and near infrared portions of the spectrum. The amount of energy that ultimately reaches the Earth's surface is dependent on the atmosphere's ability to transmit light. Figure 2.11 shows that the atmosphere is by no means transparent to all wavelengths of light. In fact, there are wavelengths at which atmospheric constituents are opaque to EM. However, as shown in the figure the atmosphere is very transparent to the visible region of the EM spectrum. The reason different colors can be distinguished is that all objects have characteristic reflectance

curves as shown in Figure 2.12. These differences can be discerned on panchromatic black and white photographs

Photography is based on photochemical changes that take place when a film is exposed to light. All film has characteristic radiometric and spectral sensitivity as well as spatial resolution.

Figure 2.11 Transmissivity of the atmosphere for wavelengths between 0.3 μm and 80 cm (Goetz & Rowan, 1981).

Figure 2.12 Typical spectral reflectance curves for vegetation, soil, and water (Swain & Davis, 1978).

2.7.2. Radiometric resolution of aerial photography

The radiometric resolution of aerial photography is related to "how a specific film - exposed and processed under specific conditions - responds to scene energy of varying intensity" (Lillesand & Kiefer, 1979). This is important for quantifying the relationship between the brightness on a photograph and the reflectance properties of the surface.

Differing illumination across the scene, atmospheric interference, and topographic effects can lead to a range of brightness for objects with the same reflectance in the ground. The type of film, exposure times (shutter speed), and filters used also have an effect on the appearance of features in a photo.

In aerial photography, there is a trade-off between contrast and radiometric range. In general, lower contrast film provides for a greater radiometric range while high contrast films offer better radiometric resolution. Ultimately, the desired film will be a compromise of radiometric resolution and range.

2.7.3. Spectral range and resolution of aerial photography

Spectral range refers to the range of wavelengths to which an imaging system is sensitive. Spectral resolution is the ability of an imaging system to distinguish between different wavelengths of light.

The total spectral range of photographic film ranges from ultraviolet to near-infrared (i.e. from less than 0.3 μm to more than 0.9 μm). To focus on particular spectral bands or to extend this range, special camera lenses must be used. Atmospheric haze is often a problem in aerial photography since the shorter wavelengths tend to be scattered by atmospheric constituents. For most panchromatic photography, a special low pass (haze) filter is used to filter these wavelengths which improves the contrast.

Panchromatic aerial photography provides perhaps the best estimate of albedo since it effectively integrates the entire visible range of the electromagnetic spectrum as shown in Figure 2.13. Higher overall reflectance is manifested as a brighter tone on an aerial photograph.

Figure 2.13 Generalized spectral sensitivities for black and white panchromatic and infrared sensitive films (Eastman Kodak Company, 1972).

2.7.4. Spatial resolution of aerial photography

Spatial resolution refers to an imaging system's ability to differentiate a feature from its surroundings. In photographic systems this is primarily a function of the size distribution of the silver halide grains in the emulsion. Larger grains reduce the resolving power but increase sensitivity to light. In this sense, spatial resolution is a trade-off between spatial resolution and film speed.

The actual ground resolution distance, *GRD*, refers to the minimum size of an object that can be resolved by a system and is represented in the following equation as

$$GRD = \frac{\text{Reciprocal of image scale}}{\text{System resolution (lines / mm) X 1000 (mm / m)}} \text{ [m]} \quad (2.21)$$

(Lillesand & Kiefer, 1979)

Image scale is dependent on the focal length of the camera and the distance between the camera and the photographed feature. For aerial photographs, the image scale can be determined from the following equation:

$$\text{Image scale} = \frac{\text{Camera focal length (m)}}{\text{Camera altitude above surface (m)}} \quad (2.22)$$

(Lillesand & Kiefer, 1979)

Image scale may not be constant throughout the entire photograph. This is dependent on the variability of the topography with respect to camera altitude. Mission parameters such as flight altitude, focal length, and film can be selected to provide sub-meter spatial resolution.

2.7.5. Temporal Resolution of aerial photography

There is no program that ensures continuous or repetitive aerial photography coverage of permafrost regions. Where more than one aerial photograph has been taken of an area, variations in mission (flight altitudes, cameras, and film) and environmental parameters (solar azimuth and elevation, humidity, time of year) mean that features in the photographs will have different tonal character and spatial resolution. In terms of digital processing, this means that some form of radiometric and spatial normalization will have to be performed prior to using the photographs for temporal studies.

2.8. Satellite imagery

Various remote sensing platforms provide data on the Earth's atmosphere and surface which is used to estimate and model the surface radiation balance (Pinker & Razgaitis, 1983a,b; Carlson et al., 1981), surface temperature (Wan & Dozier, 1989; Mannstein, 1987), and surface reflectance (Holm et al., 1989; Escadafal et al., 1989). Studies

involving surface radiation balance and temperature utilize data from weather satellites (e.g. Global Operational Environmental Satellites - GOES; National Oceanic and Atmospheric Administration's series of satellites equipped with the Advanced Very High Resolution Radiometer - NOAA AVHRR) and mission specific satellites (e.g. Heat Capacity and Mapping Mission - HCCM) with spatial resolutions ranging from several kilometres to 600 metres. The conclusions from the research show that there is a complex non-linear relationship between land surface temperatures and thermal radiation measured by satellite and that many corrections have to be applied to account for atmospheric interference and emissivity of the surface (Wan & Dozier, 1989). Other important factors such as surface roughness and surface moisture are impossible to estimate to a high level of accuracy (Mannstein, 1987). These difficulties notwithstanding, parameters such as ground level insolation have been modeled with a fair measure of success (Pinker & Razgaitis, 1983a).

Of more direct relevance to permafrost studies is the application of satellite imagery to snow cover (Dey et al., 1979), terrain classification in permafrost areas (Bélanger et al., 1990; Tamocai & Kristof, 1976), and permafrost conditions (Morrissey & Strong, 1986; Nekrasov & Petropavlovskaya, 1985; Morrissey, 1983; Schreier & Selby, 1981). With respect to snow cover, the NOAA AVHRR is able to monitor snow accumulation and ablation, with 1.1 km spatial resolution and global coverage every 12 hours. However, methods for remotely estimating snow depth and equivalent water content are still being developed (Dey et al., 1979). Terrain classification in Arctic tundra is difficult from satellite data since vegetation of all types is sparse. For example, unsupervised classification of Landsat MSS (Multi-Spectral Scanner) data for Richards Island detected only five different terrain classes including marsh, peat land, sparsely vegetated uplands, unvegetated shoreline, and partly vegetated shoreline. For the marshlands, three different moisture classes could also be distinguished (Tamocai, 1973). Where vegetation

differences are more pronounced, such as the subarctic, discontinuous permafrost region of interior Alaska, both Landsat Thematic Mapper derived vegetation maps and the TM thermal band can provide almost 80% classification accuracy in identifying frozen and unfrozen ground (Morrissey, 1986).

2.8.1. The Landsat Thematic Mapper

Satellites designed for Earth resources research are a relatively new technology. The first Landsat satellite was launched in July 1972 and was the first to have sensors designed specifically for terrestrial research. Launched in July 1982, Landsat-4 was the first to carry the Thematic Mapper (TM) sensor. Over the past decade, TM data has proven to be very useful in a variety of applications including geology, agriculture, forestry, and wildlife ecology. The limits of the TM sensor are related to its spectral, radiometric, spatial, and temporal resolution.

2.8.2. Spectral and Radiometric resolution of Landsat TM imagery

Figure 2.14 shows the spectral reflectance pattern from healthy vegetation and normalized digital numbers from the Thematic Mapper sensor. Figure 2.12 shows typical reflectance curves for water, soil, and vegetation. These figures show that the TM spectral bands were selected with a focus on vegetation. For example, a ratio of TM Band 4 and TM Band 3 can be used to better discriminate vegetation from soil and provide insight into the relative health of the vegetation. Figure 2.14 shows that the TM bands integrate the spectral reflectance and provides a simplified spectral reflectance signature but this is adequate for many forestry and agricultural applications.

The TM data quantization system has the ability to convert the measured reflectance into one of 256 values. Given current TM sensitivity, total image entropy has been found to use on the order of 64 of the 256 levels possible (Bernstein et al., 1984). This suggests that the radiometric sensitivity of the sensor is well within the limits of the related pre-

processing system. The focus of sensor development in recent years has been to increase the spectral resolution of Earth-observation sensors by decreasing the spectral bandwidth for each band and increasing the number of spectral bands.

Results from Arctic tundra regions show that current technology has difficulty discriminating subtle vegetative patterns. Future sensor technology may provide 500 spectral bands or more for the same spectral range as the six TM sensors that measure reflected electromagnetic radiation. Perhaps this will improve TM capabilities in the Arctic.

Figure 2.14 Spectral reflectance curve for healthy vegetation and corresponding normalized TM digital numbers (Mather, 1987).

2.8.3. Spatial resolution of Landsat TM imagery

The Thematic Mapper sensor has a nominal spatial resolution of 30 metres. At a scale of 1:100,000, this represents a 0.3 mm dot and is approaching the practical lower limit for map production and interpretation. For phenomena that have a much higher spatial frequency, there is some level of ambiguity in the classification of each TM sensor pixel. This, coupled with the spectral integration that was discussed in the previous section, means that ground truth is required to help verify TM image classification.

2.8.4. Temporal resolution of Landsat TM imagery

The Landsat-4 satellite maintains a sun-synchronous, near-polar orbit at an nominal altitude of 705 km. The orbit period of 98.9 minutes (14.5 orbits per day) combined with ground swath width of 185 km means that global coverage is achieved every 16 days. In practice, however, the time between cloud-free images is certainly more than every 16 days for much of the Earth. For example, for Richards Island the most recent cloud-free summer image was in July 1986. A repeat period for cloud-free images of 3 years in this area seems long relative to the operational repeat period of 16 days. However, it should be noted that vegetative changes occur very slowly in the Arctic and a relatively long repeat period may be suitable for global change studies.

2.9. Geostatistics

Geostatistics were originally applied to mining and geological problems (Matheron, 1971; Journel & Huijbregts, 1978). More recently they have been applied to modeling and deciphering spatial pattern of landform and soil properties (Oliver & Webster, 1986), designing sampling strategies for soil sampling (Trangmar & Kemp, 1989; McBratney et al., 1981a & b), examining spatial cross-correlations between soil properties (McBratney & Webster, 1983), and the relations between remotely sensed imagery and various kinds

of surface cover (Atkinson, 1987; Curran, 1987, 1988; Woodcock, 1985). No reference could be found, however, for the application of geostatistics to the depth of active layer or any related permafrost phenomena.

2.9.1. Introduction to geostatistics

Geostatistics is defined as "the application of the formalism of random functions to the reconnaissance and estimation of natural phenomena" (Matheron, 1971). It attempts to recognize both the structured and random aspects of natural phenomena. A regionalized variable is the term given to a measured quantity in space that possesses these somewhat contradictory characteristics. However, at any given point in space, the value taken by a regionalized variable is related to values within a given area. This spatial interdependence is termed autocorrelation (Cliff & Ord, 1987). Qualitatively, it is relatively simple to imagine that the autocorrelation of a given phenomena will be related to the distance between observations. The purpose of geostatistics is to quantify and describe the spatial interdependence of data. This can be done in one, two, or three dimensions. The results can be used to estimate the value of a regionalized variable at points where there is no sampling data. Kriging is the name given to the procedure that uses the information from the semivariogram to find an optimal set of weights that are used in the estimation of a regionalized variable at unsampled locations (Davis, 1986).

2.9.2. Semivariograms and quantifying spatial variability

In the most general sense, at a given sampling interval, autocorrelation or semivariance are measures of similarity or dissimilarity respectively. Lag is the term given to the sampling interval and is a parameter representing the number of intervals between samples used in computing the autocorrelation or semivariance. It is expressed as either a dimensionless number or in units of distance when plotted. Figure 2.15 shows the relationship between semivariance and autocorrelation. It shows that at a lag of 0 where exactly the same point

is sampled, the autocorrelation is 100% whereas the semivariance is 0%. As the lag increases, the spatial interdependence between the samples decreases so the autocorrelation decreases and the semivariance increases accordingly.

Figure 2.15 Relationship between semivariance and autocorrelation for a stationary regionalized variable (adapted from Davis, 1986).

The plot of semivariance versus lag is called a semivariogram, where semivariance, γ , is computed using the following equation:

$$\gamma(h) = \frac{1}{2(N-h)} \sum_{i=1}^{N-h} [z(x_i) - z(x_i + h)]^2 \quad (2.23)$$

where N = total number of samples taken along a profile

z = sample value (regionalized variable)

x_i = distance from beginning of profile to the i th sample

h = lag

The semivariance is half the average squared difference between sample values at a given lag, h . The average squared difference is divided by two since it is assumed that each sample contributes equally to the difference between any sample pair. The difference is squared for two reasons. One is that the difference between realizations of a regionalized variable at a given lag will alternate between positive and negative values. Squaring the

difference ensures that the semivariance will always be positive. The other reason is to "amplify" the differences between semivariance computations at various lags. It is assumed that the spacing between samples is constant. In situations where this is not the case, values should be interpolated to permit computation of the semivariance. The lag can be expressed in units of distance by multiplying the lag by the sampling interval.

To help visualize how semivariance is computed, Figure 2.16 shows the pairs that would be used in the computation of semivariance for the first three lags (i.e. $1 \leq h \leq 3$). In this case, the total number of samples is 7. This figure shows that the semivariance is computed using every possible data pair at a given lag. As the lag increases, the number of sample pairs decreases. Semivariance computations, like other descriptive statistics such as mean and standard deviation, require many data points to help ensure their usefulness. A minimum of 25 data pairs was used for all semivariance computations. Therefore, the maximum lag, and number semivariance values, that could be used for plotting the semivariogram was 25 less than the number of samples (i.e. $h_{max} = N - 25$).

Figure 2.16 Pairs of data points used in the computation of semivariance.

There are a number of terms used to describe the character of a semivariogram that are summarized in Figure 2.17 and defined in Table 2.6. The semivariogram shown in Figure 2.17 is different from that shown in Figure 2.15 in that the intercept with the semivariance axis is not zero. The term nugget variance is given to the intersection of an extrapolated semivariogram and the semivariance axis at a lag of zero. In theory, the semivariance must be zero at a lag of zero. Possible reasons for a non-zero nugget variance are a very erratic regionalized variable over short distances, inadequate precision of the sampling device, or a large support relative to the spatial frequency of the regionalized variable.

If the regionalized variable was highly erratic but we were able to accurately measure its amplitude at very small lags, then the result would be a very steep semivariogram at very small lags and a nugget variance of approximately zero.

For a regionalized variable with low spatial frequency, a measuring device that had a low level of accuracy might be able to provide a useful semivariogram but would probably give a non-zero nugget variance that would approximate the device's accuracy.

Support in the context of regionalized variable theory refers to the area, shape, and orientation of the surface or volume represented by each sample and is a function of the sampling device and strategy. For example, remote sensing devices often have an instantaneous field of view and resulting image pixel size that is larger than the spatial variability in the reflected or emitted radiation from the surface that they are imaging. Remote sensing devices such as aerial cameras or satellite sensors rely on spatial and spectral integration to increase signal to noise performance. The amplitude of the signals are quantized to some extent as well. The combination of this integration and quantization means that the imaging systems are unable to resolve sub-pixel spatial autocorrelation and will often have a nugget variance associated with the semivariogram. In addition, if the phenomenon being assessed is not spatially isotropic, then the orientation of the data sampling is important.

Figure 2.17 Theoretical semivariogram for data that conforms to the spherical model (adapted from Curran, 1988).

Table 2.6 Terms, symbols, and definitions used in describing the semivariogram (adapted from Curran, 1988).

Term	Symbol	Definition
Lag	h	number of intervals between sample pairs
Sili	s	maximum computed semivariance
Range	a	lag at which semivariance is 95% of sill
Nugget variance	Co	extrapolation of semivariance to $h = 0$
Spatially dependent structural variance	C	$C = s - Co$
Support		area, shape, and orientation of point, surface, or volume represented by each sample

3. Methods

This chapter will describe the field work setting from both the regional and site-level perspective. It will describe the data collected and how it was processed.

3.1. Regional overview

3.1.1. Geography and physiography

The Interior Plains of Canada encompass the region between the Canadian Shield and the western Cordillera. The Northern Interior Plains (i.e. mainland north of 60°N including Banks and Victoria islands; Vincent, 1989) comprise of a number of geologic provinces; one of which is the Arctic Coastal Plain as shown in Figure 3.1. Field work was in spring (March/April) and summer (July) 1990 on Richards Island (highlighted in Figure 3.1) located in the Arctic Coastal Plain. Since this area was near the glacial limit throughout the Quaternary, glacial advances and retreats have created a complex assemblage of physiographic subdivisions as shown in Figure 3.2 and outlined in Table 3.1.

Figure 3.1 Map showing the main geological provinces of Canada, the geographical extent of Figure 3.2, and Richards Island shaded with a cross-hatching pattern.

Figure 3.2 Physiographic subdivisions of the Tuktoyaktuk Coastlands and adjacent areas (Rampton, 1988).

A geologic province is defined as a geographic area which is unified by some or all of its geological characteristics and which can therefore be studied as a whole. In this context, the unifying geologic characteristic of the Arctic Coastal Plain is the extensive Pleistocene marine and fluvial sediments which dominate the surficial geology. These are capped with glacial tills and underlain by unconsolidated to poorly lithified Mesozoic and Cretaceous rocks that have been deposited in the Beaufort-Mackenzie Basin. Richards Island is dominated by morainal (till and associated gravel and sand), glaciofluvial (sand, minor silt, and gravel occurring as a valley train), and ice contact deposits (eskers, kames) (Rampton, 1988). The Arctic Coastal Plain has economic importance in that many of the Upper Cretaceous shales that underlie the thick Quaternary strata contain significant oil and gas deposits.

Table 3.1 Physiographic subdivisions of the Tuktoyaktuk Coastlands and adjacent areas corresponding with those shown in Figure 3.3 (Rampton, 1988).

Region (Bostock, 1970)	Divisions	Subdivisions (modified from Mackay, 1963a)
Porcupine Mountain Area (Eastern Cordilleran)	Richardson Mountains	
	Yukon Coastal Plain	(Mountain Fringe)
Arctic Coastal Plain	Mackenzie Delta	
	Tuktoyaktuk Coastlands	Mackenzie Bay Islands Big Lake Delta Plain Tununuk Low Hills Kittigazuit Low Hills Kugmallit Plain "Low Involuted Hills" West Tuk Peninsula Axis Parsons Lake Plain East Tuk Peninsula Axis McKinley Bay Coastal Plain Eskimo Lakes Pitted Plain Eskimo Lakes Fingerlands Liverpool Coastal Area Nicholson Peninsula Rufus Lake Lowland Kaglik Lake Plain Stanton Lowland Harrowby Coastal Plain
Interior Plains	Anderson Plain	North Caribou Hills South Caribou Hills Dolomite Lake Upland Campbell-Sitidgi Lowland Campbell Plain Hyndman Lake Hills Sitidgi Scarp Miner River Plain "Fluted Plains" Anderson River Upland Malloch Hill Upland

The Canadian Interior Plains contain the oldest record of the Quaternary in Canada. Stratigraphy in the Northern Interior Platform and Arctic Coastal Plain records at least three full glaciations and as many as four are evident in the Interior plains to the south. In general, the Quaternary can be thought as having three major glaciations; each less extensive than the preceding one (Fulton, 1989). The most extensive is older than 790 ka BP (i.e. prior to the Brunhes Normal Polarity Chron) in the Early Pleistocene. The second and third major glaciations occurred in the middle and Late Pleistocene respectively.

3.1.2. Topography

Virtually all of Richards Island topography is between sea level and 65 m. It has been severely modified by thermokarst, as illustrated by the shaded relief image of an area near Mason Bay shown in Figure 3.3. This image was derived from digitized 50 foot (15.24 metres) contours on a 1:50,000 topographic map sheet. Lake surfaces are seldom more than 35 metres below the adjacent topography. This topography is very typical of much of Richards Island. In the context of the physiographic divisions shown in Figure 3.2, all of the field work was concentrated in the Kittigazuit Low Hills which are "characterized by deeply inset lakes with moderately steep slopes on adjacent well drained ridges. Thin surface tills cap thick, brown, fine grained sands which form most hills and ridges" (Rampton, 1988)

Figure 3.3 Shaded relief image of a digital elevation model in the Mason Bay area highlighting the thermokarst that is prevalent throughout much of Richards Island.

3.1.3. Climate

The climate of the Richards Island area may be characterized as an Arctic coastal climate with long cold winters and short cool summers. Figure 3.4 summarizes the average annual temperature and precipitation patterns of the nearest weather station to Richards Island, approximately 40 kilometres to the east, at Tuktoyaktuk. The daily mean temperature, peaking in July at 10°C, does not go above 0°C until mid-May and returns to sub-zero values by early September. The mean total annual precipitation (computed from 1958 to 1970) is approximately 12.5 mm (Burns, 1973). Historically, most of the rainfall falls in August, with peak snowfall in October.

Figure 3.4 Temperature and precipitation data for Tuktoyaktuk adapted from Burns, 1973. (Rampton, 1988).

South of the treeline, the Mackenzie Delta is dominated by a continental subarctic climate due to the dehumidifying effect of the mountains to the west. The more northerly reaches of the Delta closer to the Beaufort Sea are characterized by an Arctic coastal climate. The confluence of these two climatic regimes causes relatively steep annual temperature and precipitation gradients. Both temperature and precipitation increase from north to south and east to west. Higher wind speeds are reported over the Beaufort Sea as compared to further south along the Mackenzie Valley. These winds help to reduce temperatures in the coastal areas in the summer and to redistribute the meager snow cover in the winter. Snow cover variations can be quite dramatic. For example, in a setting where slopes were less than 10° , a range of snow depths from less than 10 cm to more than 2 meters were found within an area of 15 square metres.

3.1.4. Vegetation

Low Arctic tundra vegetation is prevalent throughout most of the Tuktoyaktuk Coastlands as shown in Figure 3.5. The transition zone between boreal forest and low

Arctic tundra converges moving towards the Mackenzie River from the east. In this transition zone, the occurrence of trees is very dependent on topography, namely slope and orientation. It is quite common in this zone to have low Arctic tundra vegetation or bare soil on the tops of hills and north-facing slopes, while the south facing slopes are dominated by relatively large coniferous trees between 2 and 15 metres in height.

Much of Richards Island is covered by low Arctic flora including sedges, scrub birch, willows, and heaths. In poorly drained areas, tussocks often develop (Lambert, 1972). In areas that are protected from the harsh winter winds, and where moisture conditions are favourable, the scrub trees can reach heights of up to 2 metres. Where the surface is exposed to winter winds, such as the crests of hills, vegetation is very low or absent.

Figure 3.5 Vegetation of the Tuktoyaktuk Coastlands and adjacent areas adapted from Mackay, 1963; Ritchie & Hare, 1971 (Rampton, 1988).

3.2. Field work sites

Figure 3.6 shows the location of field sites on Richards Island. At sites 1 and 3 (Lousy Point and North Point), detailed frost-table measurements were made along profiles in an attempt to quantify spatial variability. At site 2, the majority of the work focused on studying the relations between the satellite imagery and terrain characteristics.

The site selection was imposed to a certain degree by the logistical support provided by the Geological Survey of Canada as part of their "Global Change" activities. Transportation of the heavy drilling and other field equipment, camping gear, and food was primarily by Zodiac inflatable dingys with outboard motors. For this reason, only areas accessible from the Mackenzie River and the Beaufort Sea were visited.

The vegetation pattern near Lousy Point (Figure 3.6) is characterized by 1 to 2 metre high dwarf birch and willow that dominate the relatively steep slopes contrasted with the very heterogeneous and much more sparse vegetative cover on the crests and tops of hills. The shrubby vegetation on the tops of hills is typically less than 1 metre in height and is interspersed with patches of bare soil, sedges, and grasses. Moving to the north, the occurrence and size of shrubs greater than 1 metre in height decreases presumably due to the increased severity of winds and colder temperatures.

All the sites were underlain by unconsolidated morainal and glaciofluvial deposits.

Figure 3.6 Map showing location of field sites on Richards Island. 1) Lousy Point, 2) Mason Bay, and 3) North Point.

Field data was acquired in the Spring (March/April) and Summer (July) of 1990 and was intended to provide information on the influence of snow, micro-topography, surface cover, and surface temperature on the depth of the frost table.

Work in the Spring was restricted mostly to Site 1: North Point, (69°39'N, 134°22'W) when measurements of snow depth and snow surface topography were made. General observations on the effect of topography on snow cover were also noted.

Work in the Summer was performed in the three areas shown in Figure 3.6: Site 1: Lousy Point (69°14'N, 134°16'W), Site 2: Mason Bay (69°33'N 134°01'W), and Site 3: North Point. Measurements of micro-topography, ground and surface temperatures, thaw depth, and slope were taken at sites 1 and 2. Surface cover types were recorded in all three study areas as well.

3.3. Field data acquisition and processing

For the field sites, three different types of data were used:

- 1) field data: measurements of thaw depth, micro- and macro-topography, surface temperature, and vegetative cover
- 2) aerial photography: panchromatic black-and-white aerial photographs were digitally scanned.
- 3) satellite imagery: sub-images from a Landsat Thematic Mapper dataset were extracted

Unfortunately, Ground Probing Radar data for these sites could not be obtained. GPR data was acquired for another area of continuous permafrost, to assess its potential.

All field notes were transferred into Microsoft's Excel spreadsheet program for plotting graphs, finding frequency distributions, and regression analysis. Comma separated ASCII files were exported from Excel as required for use with other programs. A program called GEOSTAT was written using Microsoft's Visual BASIC for Windows to compute

semivariances of the field data collected along transects and of profiles from the scanned aerial photographs. Details of the data acquisition and processing for each type of data will be discussed in the sections that follow.

3.3.1. Depth of active layer and microtopography

Figure 3.7 shows the equipment used in measuring thaw depth, slope, aspect, and microtopography. A measuring tape was attached to the aluminum rods for measuring distance along the transect. For most measurements, three aluminum rods were coupled together (as shown) and this was used as a reference for measuring micro-topography. A Brunton compass was used to measure both the slope and aspect of this "reference equipment". For every measuring point along a given transect, the distance between the reference and the ground surface as well as the distance between the ground surface and frost table were measured. In this way, the relationship between micro-topography and measured thaw depth could be investigated. Thaw depth was measured by pushing the probe into the soil until it contacted the frost table (i.e. the bottom of the active layer). The probe used in all field work was 1.25 cm in diameter, and would be considered "large" compared to other frost probes. Of the several hundred measurements taken, the method proved to be impossible in less than ten cases where there were boulders between the surface and the frost table or where the frost table was deeper than the 1.3 metre shaft of the probe.

Frost table probe

Equipment used as a constant reference for measuring micro-topography

Figure 3.7 Equipment used in measuring micro-topography and thaw depth.

3.3.2. Snow depth

Measurements of snow depth were made using a Mount Rose snow coring device. It comprises of a hollow aluminum tube that is serrated on the bottom edge, for easy penetration into hard packed snow or ice, and has graticules etched on its outer surface to measure depth. Snow surface topography (i.e. elevation, slope and aspect of the snow surface) was measured using a Brunton compass and surveyors stakes and tape.

Relatively extensive measurements of snow depth and snow surface topography were made at North Point on a slope approximately 200 m south-southeast of a Geological Survey of Canada Bore Hole (90BH1). After measuring the snow depth, the snow coring device was used to extract short cores of surficial material. This created marks in the terrain that were used as a reference for the work performed in the summer.

3.3.3. Surface and subsurface temperature

Soil temperature was measured using a TEL-TRUE soil thermometer with a range from 0° to 50°C and a resolution of $\pm 0.25^\circ\text{C}$. This is a "dial" thermometer with a 21 cm stem. The thermometer is sensitive to temperature along a 7 cm length of the stem beginning at its tip. Therefore, this type of thermometer displays the average temperature over a 7 cm thick layer of soil. The thermometer was inserted vertically into the soil so that most of the data in this report represents average temperatures from 0-7 cm and 14-21 cm below the surface. Some of the data collected early in the field work were recorded as "temperatures from 0-2 cm depth". In these cases, the thermometer was inserted at an angle so that the tip was approximately 2 cm below the surface. This method was abandoned after the Lousy Point transect due to problems with repeatability.

3.3.4. Surface cover classification

Surface cover plays an important role in the thermal regime of the ground. Classification of surface cover was made in a very general way, specifying groups such as mosses (high density/low density moss), scrub type and height (e.g. 20-50 cm Arctic birch), grass/vetch, and exposed soil (e.g. silty soil boil, sand).

3.4. Scanned aerial photography

Portions of 22.8 cm X 22.8 cm aerial photos were scanned using an Epson ES-300C scanner at a resolution of 600 dpi (dots per inch) resulting in an effective pixel size of 2.6

metres. The scanned image was stored as an 8-bit (256 level) TIFF (Tagged Image File Format) file. Using ARC/INFO, this file was converted to a BIL (Band Interleaved by Line) file and geometrically corrected to the NTS (National Topographic System) 1:50,000 scale map sheets. The map sheets used were 107C/3 E for Lousy Point and 107C/11 E and 107C/14 E for North Point. The aerial photos used were A26749-225 and A26754-200 for Lousy Point and North Point respectively.

3.5. LANDSAT Thematic Mapper imagery

The imagery used was from the Thematic Mapper sensor onboard the Landsat V satellite. The image was taken July 23, 1986. Thematic Mapper imagery may be purchased as full scenes (approx. 185 km X 185 km) or quadrants (approx. 90 km X 90 km). The imagery used was from Quadrant 4 of the scene referenced by Track 65, Frame 11.

The three computer compatible tapes (CCTs) (Tape Numbers MO1357 through MO1359) were provided by Dr. Robert Bélanger (Terrain Sciences Division of the Geological Survey of Canada). The format of the data was LGSWG (Landsat Ground Station Working Group). The EASI/PACE image analysis system from PCI Inc. was used to download the portion of the image covering Richards Island from CCT to a more generic BSQ (Band SeQuential) format. Access to the PC-based PCI system was given by Dr. Andy Rencz of the Geological Survey of Canada. Due to limited access to the PCI system, the BSQ files were backed up on floppy discs for transfer to other systems for correction, enhancement, and analysis.

The Mason Bay area was extracted from the BSQ files and imported into the RSVGA Image Analysis System from EIDETIC Digital Imaging Ltd. Geometrical correction and image enhancement of all seven Thematic Mapper bands were performed. The geometrical correction was made to the NTS 1:50,000 scale sheets 107C/11 E, 107C/10 W, 107C/6 E, and 107C/7 W. The results were exported to BSQ format and imported

into the EPPL7 Geographic Information System from Minnesota Land Management Information Corporation for analysis.

For the Lousy Point and North Point images, all seven bands for all of Richards Island were imported into the ARC/INFO GIS. To allow for fast and accurate geometrical correction, sub-images of the individual areas were extracted from the uncorrected imagery and processed.

3.6. Profile extraction and analysis from satellite imagery and aerial photography

3.6.1. Profile extraction

The term "profile" refers to pixel values from satellite imagery or aerial photography that correspond to points along a digitized line. Profiles were extracted for Lousy Point and North Point sites from Thematic Mapper bands 3 and 5, visible and near-infrared frequencies respectively, and from digitally enhanced scanned aerial photography.

The frost table probing at the Lousy Point site was done along a six metre profile with a sampling interval of 10 cm. Since the aerial photograph had a pixel size of approximately 2.5 metres, a profile of less than 3 pixels would have not been useful for statistical analysis. For this reason the profile for this site was extended for approximately 50 metres in both directions along strike with the frost table profile to evaluate whether the measured "sub-pixel" variability in the frost table was somehow reflected in the variability observed in the aerial photography.

The frost table profile at North Point was 81 metres in length. With a pixel size of approximately 2.5 metres, the number of scanned aerial photo pixels along this profile is 32 which was sufficient for statistical analysis.

3.6.2. Semivariance of profiles

Semivariance was computed for profiles by exporting only the values at each location (i.e. the z from an x,y,z coordinate) from ARC/INFO to an ASCII file. This file was then used for input into GEOSTAT. The output of GEOSTAT is a comma separated file containing lags and semivariances. This file was imported into Microsoft Excel for plotting graphs.

3.7. Ground Probing Radar (GPR) acquisition and processing

Some work has been done using GPR in permafrost terrain (Annan et al., 1975a,b, Dallimore & Davis, 1990; Pilon et al., 1992). However, the focus has been on mapping of ground ice at depths of metres to 10's of metres below the surface. Consequently, these surveys were all done at frequencies much lower than the 1000 MHz necessary to adequately map the active layer. They were also done in the winter so that the attenuation and reflection at the frost table that prevents deep penetration of the soil could be avoided. In fact, no data set could be found on short notice that was at a sufficiently high frequency and in a terrain and season that could be used to demonstrate the utility of GPR for mapping the active layer. A dataset was acquired, however, to assess the method and to provide some insight into the potential for mapping the active layer.

The data was graciously provided by Brian Moorman (Carleton University). He had used GPR on a 107 metre profile near an automatic weather station on the Fosheim Peninsula, Ellesmere Island, N.W.T. The data was collected in the summer and there was a shallow active layer present at the time. Unfortunately, there is no ground truth to provide moisture conditions or typical active layer depths since, once again, this profile was not intended for mapping the active layer. The surficial material was very poorly lithified sandstone with lenses of coal. Hydrocarbons complicate the electrical properties of the ground and can cause odd behaviour of GPR (Annan, 1992).

Five programs were written using Microsoft's QuickBASIC 4.5 to process the GPR data as provided by the pulseEKKO IV GPR system used by the Geological Survey of Canada. The names and brief description of their function is given below. These programs were used to help understand the dataset, to attempt automatic extraction of the active layer reflection from the 107 traces taken, and to provide data to the GEOSTAT program for computation of semivariance.

Table 3.3 Programs used for processing GPR data.

Program Name	Description
DT2ASC.EXE	Converts pulseEKKO IV files from binary to ASCII.
MINMAX.EXE	Finds minima and maxima above/below the given threshold in pulseEKKO IV binary files and writes the result to an ASCII file
PICKPEAK.EXE	Selects chosen MIN or MAX from pulseEKKO IV binary files and writes the result to an ASCII file.
TRCGET.EXE	Selects the specified trace from pulseEKKO IV binary files and writes the result to an ASCII file. The extension of the output file name is the trace number that was selected.
TRCPLT EXE	Displays trace 1 from specified pulseEKKO IV file. Viewing parameters and the plotted trace can be changed interactively after the first trace is displayed.

The TRCGET, TRCPLT, and MINMAX programs were used to examine a number of the 107 traces at random to select the peaks that would most likely represent the ground wave and reflection from the frost table. This involved not only selecting specific peaks along the trace but a suitable amplitude threshold as well. Peaks 2 and 3 (maxima) were selected with a threshold of 1400. The data was extracted using the PICKPEAK program in two separate passes; one for peak 2 and the other for peak 3. These were then imported into Microsoft Excel and the two-way travel time of the electromagnetic wavelet was computed and plotted. The two-way travel time was exported to an ASCII file for semivariance computations in GEOSTAT. The output of this program was then brought back into Excel for graphing.

4. Results and discussion

The objective of this section is to provide some insight into the spatial character of active layer variation and related parameters and to evaluate the utility of digitally scanned aerial photography, satellite imagery from the Landsat Thematic Mapper sensor, and ground probing radar (GPR) for active layer mapping. Figure 4.1 provides a perspective on the results and discussion that are to follow. It shows standardized values for three of the four types of data along the North Point transect (Field site 3). It is immediately obvious from this figure that the active layer, aerial photographs, and satellite data are very different in their spatial character and show little relationship to each other.

Figure 4.1 Standardized values for measured thaw depth, aerial photography, and satellite imagery (North Point site).

4.1. Direct field measurements

4.1.1. General analysis of field data

Figure 4.2 shows the results of various field measurements taken at Field Site 1 at Lousy Point. A few general observations can be made.

Measuring the average surface temperature from 0 to 2 cm was abandoned after the Lousy Point site due to problems with repeatability. However, the high spatial variability of these measurements was found to be consistent with measurements taken at North Point. If the thermal properties of the ground were spatially uniform, then the average temperature from 14 to 21 cm depth would be a subdued imitation of temperature variations at the surface. The unreliability of the surface temperature from 0 to 2 cm notwithstanding, the graph of the temperature difference (ΔT) indicates that the temperature gradient, and therefore the thermal properties of the ground in the near surface, exhibit high spatial variability. This is somewhat expected due to the spatial variability of the ground materials, vegetative cover, and moisture conditions in the near surface that were observed in the field. Field observations indicate that the spatial variability of temperature is higher than that of active-layer depth. This is exaggerated in Figure 4.2 since the scale for the active layer is compressed relative to the scale used for plotting temperature. The differences in spatial variability suggests that the thaw depth will be poorly correlated with the temperature data as illustrated in Figure 4.3. There are several reasons for this:

- thaw depth is related to both the spatially integrated temperature at the surface and the thermal properties of the ground for several centimetres around the measuring point.
- the temperature measurement is instantaneous whereas the depth of thaw is related to temperature patterns over a longer time scale

- local factors, such as vegetation cover, affect the annual surface temperature regime (Smith, 1975) and are presumed to affect instantaneous temperature as well.
- many of the surface temperature maxima are associated with moss, which has an extremely low thermal diffusivity at the surface, and are thus correlated with thaw depth minima.
- effect of the latent heat of vapourization. If the soil contains moisture, energy will be dissipated in the evaporation of sub-surface water. This is one of the more difficult parameters to account for in estimating the depth of thaw (Brown, 1964).

The average temperature from 14 to 21 cm below the surface exhibit lower spatial variability and a more reasonable correlation with thaw depth.

Given the difference in thermal properties of the ground, the expectation is that the frost table will not be a subdued reflection of the micropography at the surface. The surface and frost table profiles in Figure 4.2 and Figure 4.4 shows that this is the case. The distance between these two profiles is the thaw depth (depth of the active layer). Figure 4.4 also highlights the effect on the surrounding frost table of an anomalous feature such as a soil boil. In this case, the proximity of the other surface cover types to the soil boil leads to a thaw-depth measurement that exceeds the average for each cover type. Such proximity effects apply to all thaw depth observations and increase the measured variability in the active layer for a given surface cover. Figure 4.5 shows that the total variability in active-layer depth ranged from 10 cm. for moss (i.e. $21 \text{ cm} \pm 5 \text{ cm}$) to 36 cm. for small alder and birch less than 20 cm in height (i.e. $46 \pm 18 \text{ cm}$). All of these measurements were made within a 10 x 10 metre area containing a variety of surface cover types and moisture conditions, resulting in thaw depths ranging from 16 to 80 cm.

Figure 4.2 Temperature, microtopography, frost table, and thaw depth data derived from field observations at Lousy Point (Field Site 1).

Figure 4.3 Correlation between average temperatures, measured at 0-2 cm and 14 - 21 cm depth, and the thaw depth.

Figure 4.4 Microtopography and frost table profiles illustrating sensitivity to surface cover and ground thermal properties.

Figure 4.5 Differences in thaw depth due to variations in surface cover. Circles represent the mean value based on all data shown with an X.

Since the frost table is not necessarily a subdued reflection of surface microtopography, thaw depth measurements were made relative to a standard reference. Ignoring the effects of microtopography leads to misleading impressions of the frost table and its spatial variability. To demonstrate this, the data from Figure 4.4 has been replotted in Figure 4.6 to show the frost table and thaw depth. The frost table was computed by adding the distance between reference apparatus and the ground surface to the measured thaw depth. This Figure shows that the rise in the microtopography due to the tussock (between 210 and 250 cm from the beginning of the transect) is not reflected proportionately in the frost table, and the difference between the measured thaw depth and the depth of the frost table relative to a standard reference is approximately 14 cm. The microtopography and proximity effects on the frost table illustrated in the above figures show that thaw depth measured at a point is the result of three-dimensional conditions and processes and field procedures should attempt to account for them.

Figure 4.6 Measured thaw depth and frost table profile. Shows potential errors in measuring thaw depth without accounting for microtopography.

It has already been noted that the most of the topography of Richards Island is between sea level and 65 m. Therefore, altitude effects that are often incorporated into permafrost modeling (Evans et al., 1989) can be ignored here. The topographic characteristics of slope and aspect, however, are recognized as major factors in controlling the type of vegetative cover (Davis, et al., 1990), total insolation (Oke, 1987), and snow cover (Smith, 1975; Brown, 1978). It is not surprising, therefore, that virtually all permafrost models include some reference to slope and aspect (Peddle, 1992; Lunardini, 1978). Figure 4.7 shows the effect of slope and aspect on thaw depth for two different surface cover types. The Figure clearly shows that variations in thaw depth within each feature is far greater than the distinct effect of slope, aspect, and topographic position. This is not the case in areas of discontinuous permafrost where topographic effects are more noticeable (Peddle, 1992). Therefore, models and methods that are applied in areas of discontinuous permafrost where vegetation and topography play a greater role in permafrost distribution are probably not effective in the higher latitudes where the vegetation is dominantly low Arctic tundra and the lapse rate due to elevation changes is insignificant.

KEY:

- 1 = near bottom of north-facing slope
- 2 = near top of north-facing slope
- 3 = on top of hill (relatively level ground)
- 4 = middle of south-facing slope

Figure 4.7 Variations in thaw depth, for grass tussocks and moss/heather filled depressions, with changes in topographic position, slope, and aspect.

Snow cover has been recognized as one of the major factors in determining the ground thermal regime (Nicholson & Granberg, 1973; Smith, 1975). Figure 4.8 shows the relationship between snow depth and thaw depth for the collinear summer and winter transects at North Point. This graph shows that thaw depth is best developed for snow depths less than 0.5 m and greater than 1.25 m. Snow depths in the range of 0.75 m seem to limit the depth of thaw. The polynomial regression equation and graph are presented to highlight the trend in the data rather than to suggest a relationship to predict thaw depth from snow depths. The envelope represents the possible range of values due to other factors for a given snow depth at this site. It is apparent that there is a substantial degree of ambiguity in the relationship of the active layer and snow depth. Since snow insulates the ground from the extreme cold of the winter months, it can be assumed that the ground does not get as cold when the snow depth is greater than 1.25 m. Therefore, less energy is required to bring the ground to melting temperature so, for the same energy input, the active layer will be more developed. On the other hand, snow cover can delay the onset of the thaw cycle by shielding the ground from early Spring sunshine. If there is little or no snow cover, the ground will get very cold in the winter but will experience a longer thaw season. Other studies have shown that the timing of initial snow accumulation is an important factor that affects annual ground surface temperature (Goodrich, 1982) and, presumably, the development of the active layer during the thaw season. The influence of these factors is reflected in Figure 4.8. For this reason, a relationship between snow depth and active layer development is not likely to be easily implemented for mapping the active layer.

Figure 4.8 Thaw depth versus snow depth for the summer and winter transects at North Point.

4.1.2. Geostatistical analysis of field data

Now that the general character of thaw depth, surface temperature, microtopography, surface cover, and the ground thermal regime have been reviewed, the spatial variability of these variables and data provided by scanned aerial photography, satellite imagery, and ground probing radar will be examined.

Figure 4.9 shows the character of the ground surface and frost table at Lousy Point. The values represent the distance below the standard reference. The thaw depth is the difference between these two profiles. The first 150 cm of this transect are different in character from the remaining 450 cm due to the effect of a small pond on the ground surface and frost table. Figure 4.10 shows the semivariance of the thaw depth with and without the effect of the pond. It is clear that for lags longer than 50 cm, the pond causes a substantial difference in the spatial variability as estimated by the semivariogram. For this reason, all other data sets involving the Lousy Point transect were computed for measurements between 150 and 600 cm along the transect.

The semivariance, as defined in Chapter 2, is the average squared difference between data values at a given separation or lag. To associate this with common measurement systems, Figure 4.11 is a plot of the same data with semi-standard deviation plotted against lag. This was computed as the average absolute difference of measurements at different lags. If the graphs are projected to zero lag, the intersection with the semi-standard deviation axis is approximately 1.5 cm. Theoretically, this means that no matter how small the sampling interval is, there is always going to be a ± 1.5 cm variation in the measured thaw depth for this particular transect. For the first 50 cm, the graphs increase in a linear fashion and this region is called the range. This means that the thaw depth measured at any point along this transect is autocorrelated with other thaw depth measurements within 50 cm. Beyond 50 cm, the slope of the graph without the effect of the pond diminishes approaching a maximum called the sill. In this case, the sill appears to be between 10 and 12 cm. This means that for a sample spacing or lag of more than 50 cm will result in a measurement variation of ± 10 cm to ± 12 cm (i.e. a total variation of 20 to 24 cm). Comparing Figures 4.10 and 4.11 shows that the difference in the spatial variability are better differentiated with semivariance rather than semi-standard deviation. For this reason, the remainder of the variograms will be presented with semivariance.

Figure 4.9 Spatial character of the ground surface and frost table at Lousy Point.

Figure 4.10 Semivariance of thaw depth at Lousy Point site with and without the effect of a small pond.

Figure 4.11 Semi-standard deviation for the same data plotted in Figure 4.10.

Measured thaw depth depends on both the variability of the ground surface and the frost table. Figure 4.12 and 4.13 show the distance from a standard reference to the ground surface, the measured thaw depth, and the distance from a standard reference to the frost table for the Lousy Point and North Point transects respectively. The spatial variability in the thaw depth includes contributions from the ground surface and frost table. The degree to which the variations in the frost table correspond to variations in the ground surface will determine the variability in the thaw depth. It has already been shown that thermal properties of the ground are highly variable and this decreases the correlation between the ground surface and the frost table. The thaw depth, therefore, should exhibit higher spatial variability than either the ground surface or frost table. Both Figures 4.12 and 4.13 seem to indicate that this is the case for Lousy Point and North Point.

For comparison, it is important to note that the total transect lengths are 6 and 90 metres with sample spacing of 10 and 50 cm for Lousy Point and North Point respectively. The semivariograms for this data are presented in Figures 4.14 and 4.15. Both show that the variation in the thaw depth is greater than either the ground surface or frost table for lags of up to 1 meter. For Lousy Point, Figure 4.14 shows that the frost table exhibits higher variability at longer lags than either the ground surface or thaw depth. At first, this seems to be contrary to expectations. However, an undulating character in the frost table at low spatial frequency is observed that is not as clearly evident in either of the other graphs. For North Point, Figure 4.15 shows that, in general, the frost table and ground surface are less spatially variable than the thaw depth. In the first 5 metres, there is a range of 2 or 3 metres within which measurements seem to be autocorrelated. From 3 to 5 metres there is an intermediate sill for the thaw depth of 50 cm^2 ($\approx \pm 7 \text{ cm}$). In the context of the entire semivariogram, however, there seems to be a range of approximately 25 metres for the thaw depth. That is, thaw depth seems to be autocorrelated for lags of up to 25 metres. It is highly improbable that the ground thermal regime at a given point along the transect is

2 of/de 2

autocorrelated with another point 25 metres away in the context of heat transfer processes. However, it is likely that the spatial pattern in the thaw depth is related to a similar sets of conditions related to ground materials, surface cover, and other factors.

Figure 4.12 Distance from the ground surface and frost table to the standard reference and the measured thaw depth for the Lousy Point site.

Figure 4.13 Distance from the ground surface and frost table to the standard reference and the measured thaw depth for the North Point site.

Instead of remaining at the apparent sill at a lag of 25 metres, the thaw depth semivariance begins to decline in a way similar to the way it increased for lags less than 25 metres. Summertime active layer measurements started at approximately 31 metres along the winter transect as shown in Figure 4.13. Between 55 and 80 metres along the transect, there is a change in both the amplitude and spatial character of the thaw depth. The last 40 metres seem to have a similar pattern to the first 25. This spatial pattern is highlighted by the semivariogram in Figure 4.15. The field data shown in Figure 4.1 showed that the active layer from 25 to 45 metres was not as well developed as it was from 0-25 metres and 45-80 metres. In the field, these three sections were characterized by different surface covers. From 0 to 25 metres, the transect was covered mostly by thick grasses. From 25 to 45 metres, the ground was covered by organic matter and sparse vegetation. From 45 to 85 metres, the vegetation included vetch, dwarf birch, and willow all less than 50 cm in height.

Semivariograms are robust tools for highlighting and quantifying spatial variation. It is apparent that there are differences in the amplitude and frequency of the three curves in Figure 4.15. Of particular interest is the ground surface semivariance. Two characteristics are apparent from this curve. One is that there seems to be very little autocorrelation in the ground surface beyond 2 or 3 metres. This confirms field observations that microtopographic features along this transect were typically less than 3 metres in extent. The other characteristic is that the ground surface seems to exhibit periodic changes in spatial variability and this pattern recurs at 6 metre intervals. This pattern appears to be a result of the method used in measuring the ground surface. The standard reference was 3 metres in length and was moved end-on-end up the gently sloping transect. The inclination angle of the standard reference changed with each move by as much as 6° and this affected the measured distance between the reference and the microtopography. This 6 metre cycling is not at all apparent in the original ground surface

shown in Figure 4.13 from which the semivariance was computed. This exemplifies how semivariograms help to decipher and quantify spatial variability and patterns.

Figure 4.14 Semivariogram of data presented in Figure 4.12 at Lousy Point site.

Figure 4.15 Semivariogram of data presented in Figure 4.13 at North Point site.

The semivariograms for surface temperatures at Lousy Point and North Point shown in Figures 4.16 and 4.17 respectively show that there may be autocorrelation at lags of 30 cm or less. As expected, the semivariance of the average temperature at 14 - 21 cm depth is much less than it is at 0 - 2 cm or 0 - 7 cm. Figure 4.3 showed that thaw depth and surface temperature are not highly correlated. It is somewhat surprising, therefore, to see a peak in the semivariance of the surface temperature at lags of approximately 25 metres as was the case with the thaw depth in Figure 4.15. This suggests that surface temperature variability is mirrored to some extent by thaw depth variations for the North Point transect.

Figure 4.18 a) shows that, sampling interval notwithstanding, there is a dramatic difference in the spatial character of the first six metres of the North Point transect and the 4.5 metres of the Lousy Point transect that are not affected by the small pond. This point is punctuated by the semivariograms of the Lousy Point and North Point transects in Figure 4.18 b).

The semivariograms of surface temperature and thaw depth (Figure 4.15 and 4.17 respectively) show that there is some relationship between the spatial variability of these parameters. A question that needs to be addressed is whether the variability in the active layer can be inferred from other sources such as aerial photographs.

Figure 4.16 Semivariogram of average surface temperatures (0 - 2 cm and 14 - 21 cm) at Lousy Point site.

Figure 4.17 Semivariogram of average surface temperatures (0 - 7 cm and 14 - 21 cm) at North Point site.

(a)

(b)

Figure 4.18 (a) Shows thaw depth vs distance for first 6 metres of North Point transect and the last 4.5 metres of the Lousy Point transect. (b) Semivariograms of the data shown in (a).

4.2. Scanned aerial photographs

Since the pixel size of a scanned aerial photograph is on the order of 2 metres, a minimum transect length of 50 metres is required to compute the semivariogram (recall that the criterion for computing the semivariance in this thesis is a minimum of 25 sample points). For North Point, a profile of the scanned aerial photograph was taken that corresponded exactly with the thaw depth profile. For Lousy Point, a profile was taken for 50 metres on either side of the small pond with the same azimuth as the thaw depth profile

Figure 4.19 shows that there is no correlation between thaw depth and the pixel value from a scanned aerial photograph at the North Point site. Since the aerial photo provides a surrogate for albedo, it is inferred that the thaw depth is not strongly correlated with albedo. As an example of this from the field, areas of exposed sand and gravel had thaw depths that often exceeded 1 metre yet the albedo was very high. According to the theory developed in Chapter 2, high albedo means that less energy is available for heating the soil and should therefore limit depth of thaw. Clearly this is not the case, due primarily to the thermal properties of sand and gravel (high thermal conductivity) and the absence of evaporation at the surface (high infiltration rates)

Figure 4.19 Scatter diagram showing the lack of a relationship between thaw depth and pixel value from scanned aerial photograph.

Figure 4.20 shows that there is a marked difference in the semivariograms of the two scanned aerial photo profiles. The sill for Lousy Point is approximately 180 ($\approx \pm 6$ to ± 7 in terms of the pixel value) compared to 80 ($\approx \pm 4$ to ± 5) for North Point. This suggests that the Lousy Point profile has a higher level of variability than North Point. Comparing semivariograms with lags up to 3 metres in Figure 4.18 b) with lags less than 12 metres in Figure 4.20 shows similar features. In particular, the slope of the semivariogram is much steeper for Lousy Point. In this way, even though Figure 4.1 showed that there is a significant difference in the spatial variability of the frost table as compared to the aerial photography, perhaps the scanned aerial photograph provides some insight into the relative spatial variability of the frost table at the two sites.

Figure 4.20 Semivariograms of data extracted from profiles of scanned aerial photographs at Lousy Point and North Point.

4.3. LANDSAT Thematic Mapper satellite imagery

Figure 4.21 shows a portion of the Mason Bay field site from the perspective of a scanned aerial photograph and Landsat Thematic Mapper sensor. The satellite image is the first component resulting from a principle components analysis of the three visible bands of the Thematic Mapper. As such, it is a surrogate for albedo as is the scanned aerial photograph. There are clear correlations between these two images even though they were taken years apart and at different times of day. However, the pixel size of the Thematic Mapper is 30 metres. In terms of computing the semivariance, any reasonable number of pixels would result in a profile approaching 1 km in length. In this environment, a 1 km transect spans many surface cover types and ground thermal regimes. For this reason, different approaches are required. Two types of analysis were attempted. The first involved selecting two very different thaw depth regimes and evaluating whether the satellite imagery could differentiate between them. The second compared the value of surface illumination computed from a digital elevation model with the surface temperature as measured by the Thematic Mapper thermal band (Band 6) to assess whether macrotopography can be inferred from satellite imagery.

In the first analysis, two very different environments with distinct thaw depth ranges were selected from the aerial photograph. These were elevated areas with little or no vegetation or moisture and areas with patterned ground and good vegetative cover. These areas have thaw depths typically greater than 75 cm and from 25 to 60 cm respectively. The areas were digitized from the aerial photograph, and the polygons rasterized and used as site masks for the satellite imagery. The mean values within these masks were computed for each of the raw thematic mapper bands. Figure 4.22 shows the difference in the standardized means of the different TM bands. These results show that TM bands 1 to 4 and band 6 can differentiate these environments to some extent. The difference in means is relatively small and would be expected to increase as the differences in vegetative cover

increase, as it would in regions below the tree line (Graham, 1983). This preliminary result would have to be verified with much more ground truth and analysis but does show some promise.

Black and white panchromatic aerial photography

Landsat Thematic Mapper image - 1st principle component of visible TM bands

Figure 4.21 Aerial photograph and satellite imagery at a scale of approximately 1:30,000 of the Mason Bay site highlighting differences in spatial and radiometric resolution.

Figure 4.22 Difference in standardized means for all seven Thematic Mapper bands in two distinct thaw depth environments.

In the second analysis, a digital elevation model (DEM) was created by digitizing contours from a 1:50,000 scale topographic map with a 15 m (50 ft) contour interval and interpolating these contours into a continuous surface using an Inverse Distance Weighted linear interpolation method. Since the time the satellite passed over the area is known, then so too is the solar azimuth and elevation (148° and 52° respectively). The local surface illumination was computed from the DEM. It is computed as the angle between the Sun and the normal to a 3 x 3 pixel surface on the DEM. The mean and standard deviation of the illumination for each value in the thermal band of the Thematic Mapper was computed and the result of this analysis is shown in Figure 4.23. The figure shows a weak but positive correlation between computed illumination and thermal emittance.

Figure 4.23 Relationship between illumination computed from a digital elevation model and the value from the Thematic Mapper Band 6 (thermal band).

Although the relationship between surface temperature and thaw depth was shown previously to be unclear, the thermal band of the Thematic Mapper appears to be sensitive to topographic effects and could be used potentially to provide further confirmation of a classification based on more detailed information. However, in Figure 4.24, which shows frequency distributions of both the Lousy Point and North Point transects and the corresponding pixel values from the Thematic Mapper's thermal band, the pixel values appear to be reversed relative to the thaw depths. That is, the warmer of the two areas has a less developed active layer. This shows that the application of the thermal band to even broadly defining thaw depth conditions must be done carefully.

Figure 4.24 Frequency distribution of thaw depths at Lousy Point and North Point with value from Thematic Mapper Band 6 (thermal band) for comparison.

4.4. Ground probing radar

GPR could be used effectively to map the active layer due to the electrical and electromagnetic properties of the active layer as compared to the underlying permafrost. A frequency of 1000 MHz was suggested as a possible optimum for mapping the active layer when it is less than 1 metre in depth (Chapter 2). However, no data was readily available at the time of writing, at a sufficiently high frequency to analyze the active layer. An attempt was made, therefore, to define a methodology that could be used to automatically select the frost table reflection from GPR traces.

Figure 4.25 shows a typical trace from a GPR system. The first positive peak represents the air wave. In this case, the electromagnetic pulse travels directly from the transmit to the receive antenna though the air at the speed of light. The second positive peak

represents the ground wave and is the pulse that travels along the surface. The speed and amplitude of this pulse is related to the dielectric properties of the surface which in turn is dependent on the surface moisture conditions. The third positive peak is substantially attenuated and represents the reflection from the frost table. In this case, the return from the frost table appears to be corrupted compared to the fourth and fifth peaks. This kind of corruption can be caused by either multiple reflections or constructive/destructive interference. This type of peak causes trouble in automatically selecting the correct local maximum to represent the reflection from the frost table.

Figure 4.25 Typical trace from a Ground Probing Radar survey in an area of continuous permafrost.

Figure 4.26 shows the two way travel time as determined by subtracting the arrival times of the ground wave and the third peak which is presumed to be the base of the active layer. This GPR profile was taken near an automatic weather station in the Hot Weather Creek area, Ellesmere Island, where active layer thickness is very shallow. Most of the reasonable two-way travel times range from 8 to 16 ns. Assuming a typical propagation velocity of 0.08 m/ns, this translates into an active layer depth ranging from 32 to 64 cm. Figure 4.27 is a semivariogram of the data presented in Figure 4.26. Of particular interest is the nugget variance of approximately 4 to 5 ns which translates into approximately 20 cm. Some of this high spatial variability at small lags is likely an indication that the 200 MHz centre frequency used in this profile was not high enough. A higher frequency such as 500 or 1000 MHz would surely result in a higher degree of accuracy and precision. Nevertheless, there is a discernible pattern in the active layer portrayed by the GPR profile. It shows, like the North Point site, a periodic variation in the semivariance. The range extends to a lag of approximately 30 m. The semivariogram suggests that there is a similar variation in the active layer on either side of a zone that has a spatial variability that is quite different from the rest of the profile.

The results show some promise for the automatic extraction of frost table information from the GPR data. This is an important finding given that GPR surveys can be much faster than frost table probing and can provide direct information about the frost table. However, this method does require ground truth for moisture conditions, soil texture, and thaw depth. Microtopography must also be considered to accurately map the frost table as discussed earlier (Section 4.1)

Figure 4.26 Two-way travel time to frost table as determined by taking the difference between the ground wave and the next local maxima.

Figure 4.27 Semivariogram of the data presented in Figure 4.26.

5. *Conclusions*

The results and discussion have provided some insight into the spatial character of the active layer and related conditions in an area of continuous permafrost with low Arctic tundra vegetation and unconsolidated Quaternary deposits. It must be stressed that these results may not apply in other permafrost areas where the climate, vegetation, ground materials, and topography are different. However, the methodology utilized here could be used in designing an appropriate field sampling strategy in other areas. Indeed, geostatistics should be utilized more often as a tool to help in the measurement, analysis, and description of natural phenomena.

5.1. Direct and indirect measurements of the active layer and related phenomena

Direct measurements and observations were made of active layer depth, surface temperature, microtopography, and surface cover type. Indirect measurements included thaw-depth estimation using Ground Probing Radar, pixel values from scanned aerial photography as a surrogate for albedo, and terrain classification from satellite imagery.

Surface temperatures were found to be poorly correlated with thaw depth. For example, high surface temperatures were often associated with low density mosses that had low thermal diffusivity at the surface. A few centimetres below the surface, however, moisture content and latent heat of such mosses was quite high. The combination of low energy input due to the low diffusivity at the surface and high latent heat at depth, resulted in thaw depths that were typically less than 20 cm. Similar surface temperatures were measured for sandy environments but the thaw depths were typically greater than 80 cm. The major difference in this case is that of the thermal diffusivity of sand as compared to moss. Since similar surface temperatures seem to be related to very different active layer depths, it could be assumed remote sensing of thermal emittance would not be an effective

tool for mapping the active layer. However, the thermal band of the Landsat Thematic Mapper sensor, with a spatial resolution of 120 metres, was capable of discriminating between two different active layer environments (i.e. dry uplands with thaw depths >75 cm and wet, flat patterned ground areas with thaw depths ranging from 25 to 60 cm). Spatial resolution of active layer classification depends on the infrared scanner; there are airborne infrared scanners that can provide spatial resolution on the order of a few metres or less. However, given the lack of correlation at a given point, surface temperatures are probably only useful for distinguishing between very different active layer environments. The vertical resolution of active layer measurements using surface temperature as a guide is limited to broad classification of active layer depths into perhaps no more than three classes (e.g. <25 cm, 25 to 60 cm, and >75 cm in this case). Classification accuracy, however, would depend on the provision of ancillary information such as the thermal properties of the ground, especially bulk density and moisture conditions.

The frost table is not a subdued reflection of microtopography at the surface due to variations in ground thermal properties. Therefore, for very accurate measurement of the active layer, microtopography must be taken into account. Altimetry using remote sensors has improved to the point where accuracy of a few meters and precision of a few centimetres is technically possible. However, the sub-meter spatial resolution required would make data collection and processing prohibitive. Perhaps the best alternative would be the mapping of a surface roughness factor from another sensor, such as Synthetic Aperture Radar, which would provide an estimate of the possible error in active layer measurement due to microtopographic effects.

Variations in the thaw depth for a given surface cover type were found to be large (i.e. up to $\pm 50\%$). This means that an accurate surface cover classification would not be adequate for discriminating thaw depth. These variations are due to the surface cover types, moisture conditions, and ground thermal conditions of adjacent areas (i.e. within 2

metres). In order to account for these "proximity effects", it is necessary to determine the magnitude and nature of these interrelationships.

Pixel values from scanned aerial photography, which are related to albedo, have little relationship to thaw depth. In theory, the surface temperature should decrease with higher albedos and the active layer should be less well developed since less energy is available to be transferred into the ground. This was found not to be the case mostly due to counteracting effects, such as the thermal properties of the ground. For example, sandy areas with a high albedo often had the highest surface and subsurface temperatures as well as the deepest thaw depths.

Pixel values from satellite imagery (i.e. Landsat Thematic Mapper) show some promise in discriminating different thaw depth environments but are unable to resolve the inherent spatial variability in the active layer. Results show that the active layer exhibits very high spatial variability. Within a 10 square meter area, it is possible to find active layer depths which span almost the full range of possible values for most of Richards Island. It is obvious that a representative active layer depth for any given 30 metre Thematic Mapper pixel would be inherently inaccurate.

Ground Probing Radar (GPR) can be used to measure the depth of active layer directly but must be used in conjunction with substantial ground truth to quantify important factors such as moisture content and soil constituents. Automatic selection and computation of the time difference between the ground wave and first strong sub-surface reflection shows promise for rapid extraction of active layer information from GPR surveys. Results from a 200 MHz GPR system show that there is a 4 to 5 ns (± 20 cm) two-way travel time envelope that is related to system parameters, especially the low centre frequency and long sample interval (0.8 ns). For more accurate active layer measurement, a centre frequency of 500 - 1000 MHz should be used with sample rates less than 0.2 ns in areas of continuous permafrost where thaw depths are typically less than 80 cm. There is currently

no data at this frequency in areas of continuous permafrost specifically focused on measuring active layer depth.

5.2. Geostatistics applied to measurements of active layer depth and related phenomena

Semivariograms were found to be useful in quantifying spatial variations in active layer depth and related phenomena as provided by frost table probes, aerial photography, and GPR. Semivariograms are robust tools for deciphering pattern from spatial data that appear to be random. However, they are also sensitive to spatial anomalies which can dramatically alter their character. For example, it was found that the semivariograms of the active layer with and without the effect of a small pond were very different. Consideration of such anomalies must be accounted for in the use of semivariograms in quantifying the spatial variability of any natural phenomena.

As noted in the previous section, the frost table is not a subdued reflection of the surface microtopography. The active layer, therefore, will include characteristics of both of these parameters. Semivariograms suggest that, in general, the spatial variability of the active layer is higher than either the surface or the frost table. This is a sensible assertion since the active layer is the distance between the surface and the frost table and, as such, would combine the spatial variability of both these components.

Semivariograms of active layer depth show spatial autocorrelation for lags of up to 25 metres. Autocorrelation at this scale is due to surface cover, changes in soil texture, or variations in water content. A much higher degree of autocorrelation was found for lags of 2 metres or less. The active layer at this scale is related by thermal properties of the ground, surface cover type, and subsurface heat transfer.

Surface temperature showed very little spatial autocorrelation. That is, temperatures at any given point showed little correlation with neighbouring points. However, for the

North Point site, a slight peak in the semivariance of the temperature at lags approaching 25 metres corresponded with the peak in the active layer semivariogram. This suggests that both the surface temperature and thaw depth at 25 metres from the beginning of the transect are different from the beginning and end portions of the transect. Both active layer and surface temperature reflected a change related to surface cover and, presumably, other related parameters. The spatial variability in ground temperature decreases with depth. The semivariograms highlight this marked decrease in the spatial variability of the average temperature at 14-21 cm depth compared to the upper 7 or 2 cm.

The pixel values from scanned aerial photographs were found to be virtually uncorrelated with active layer depth. However, the semivariograms of the active layer and scanned pixel values had similar slopes when comparing the two sites. That is, the site that exhibited higher spatial variability in active layer depth also had higher spatial variability in the pixel values from scanned aerial photographs. The variability of the active layer cannot be resolved by the 2.6 metre pixels of the scanned aerial photograph. However, it appears that the spatial variability portrayed by the aerial photograph could be used to infer higher spatial variability in the active layer. This observation would have to be confirmed by extensive field work.

Semivariograms of the GPR data show that the nugget variance is on the order of 5 ns^2 which is very similar to the envelope of values mentioned in the previous section. This is equivalent to a system accuracy of 9 cm (i.e. $\pm 4.5 \text{ cm}$) as computed from the following relation using a speed of propagation of 8 cm/ns:

$$\text{Accuracy (cm)} = \frac{\sqrt{\text{semivariance of two-way travel time (ns)}^2}}{2} \times \text{speed of propagation (cm/ns)}$$

The frost table probe provided nugget variances of approximately 10 to 20 cm^2 (i.e. from ± 1.6 to $\pm 2.2 \text{ cm}$). Increasing the sampling centre frequency and reducing the sampling interval, as suggested, would result in a GPR system with accuracy approaching that of

frost table probes. However, recall from Chapter 2 that the measured two-way travel time is a function of the dielectric properties of the material through which the electromagnetic wavelet passes. This means that the accuracy ultimately depends on the estimate of dielectric properties of the active layer. This may prove to be a difficult task given the natural variability in soil moisture conditions.

It is clear from the results of this thesis that the active layer is a virtually unmappable phenomenon at reasonable scales of 1:50,000 or smaller. The best possible product would combine general active layer classification with an estimation of variability for each class. Results show that the spatial variability of the active layer is not constant and could be used as ancillary information in describing the active layer.

5.3. Recommendations for further work

Based on the results of this work, a number of recommendations can be made:

- The application of geostatistics, and the semivariogram in particular, has been shown to be useful in quantifying the spatial variability of the active layer and related phenomena. More work could be done to explore the relationships between different terrain types and the spatial variability of the active layer. That is, work towards a map of semi-variance that would compliment a more generalized classification of the active layer.
- Spatial variability of pixel values from aerial photographs and frost table probe measurements of the active layer seem to be related. This finding is based on very limited field data. Further work could be done to verify and explore this relationship.
- Satellite imagery shows some promise in being able to distinguish different active layer environments (e.g. unvegetated uplands and wetlands with patterned ground) as well as macrotopography. Even though it cannot be used to accurately estimate the depth of active layer, it has been shown that the spatial variability of the active layer

seems to be related to phenomena operating at scales that can be distinguished remotely. In this way, satellite imagery may be able to help describe the spatial variability of the active layer. This possibility should be examined given the synoptic and repetitive view of Earth-observation satellites.

- GPR surveys could improve the coverage of direct measurement of the active layer using frost table probes. Automatic extraction of the ground wave and first strong subsurface reflection, and computation of the two-way travel time, has been shown to be an effective way of rapidly mapping the active layer. Using frost table probe measurements as ground truth, more data could be collected using GPR. Surveys using a GPR system with a higher centre frequency and faster sampling rate should be attempted in an area with good ground truth (i.e. frost table probe measurements), high quality aerial photographs, and satellite imagery.

5.4. Relevance and possibility of mapping the active layer

With a view to creating a map with a scale of 1:50,000 or smaller, the spatial variability of the active layer has been estimated using geostatistical techniques. The findings show that the spatial variability of the active layer is very high (i.e. sub-metre) and this has consequences in designing appropriate sampling strategies and the inapplicability of naive interpolation of sparse sampling schemes. A map of active layer, at a nominal scale of 1:50,000, to an accuracy of $\pm 10\%$ is not possible in many areas where variations of $\pm 50\%$ for a given surface cover type within a 10 metre area have been observed. The best possible active layer map would combine a general active layer classification and estimated spatial variability. Results of this work indicate that this might be a feasible goal. Semivariograms have been shown to be a useful tool in quantifying spatial variability and represent a prudent starting point for designing sampling strategies and evaluating the possibility of mapping many natural phenomena.

The utility of an active layer map, or series of maps, to global warming or impact assessment studies is not known. The recommendations for further work described in the previous section would require substantial resources. Before any effort is expended, the possible uses for an active layer map should be evaluated.

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7. *Appendices*

7.1. On-line search for geostatistics applied to permafrost

The search was performed using the DIALOG Online system through the Carleton University Library with keywords and logical associations. The nature of the search was as follows:

Provide all references from Physical Geography Abstracts that mention the following in their title or keywords:

permafrost OR frozen OR frozen ground OR tundra OR active OR layer
OR active layer OR frost OR action OR frost action OR thaw OR depth
OR thaw depth

AND

kriging OR regionalized OR variable OR theory OR regionalized variable
theory OR semivariogram OR autocorrelation OR fractal OR geostatistics

This search produced one reference:

Outcalt, S.I., Hinkel, K.M., & Nelson, F.E. (1992) "Spectral signature of coupled flow in the refreezing active layer, northern Alaska." *Physical Geography*, vol. 13, No. 3, pp. 273-284.

This article presented the results from a time series analysis of temperature logs including variance spectra. It did not refer to the spatial dimension of the active layer or related phenomena.

7.2. Field data from Lousy Point field site

Location: Lousy Point, 50 m south of 90FT5
 Date: 18/7/90
 Time: 1:00 pm-2:45 pm
 Air temp.: 20.5°C

Bearing: 86° (assume 38° mag. declination)
 Date: 19/7/90
 Time: 10:30 am-11:40 am
 Air temp.: 18.1°C

X (cm)	Temperature (°C)		ΔT (°C)	Zslc (cm)	F.T. (cm)	X (cm)	Temperature (°C)		ΔT (°C)	Zslc (cm)	F.T. (cm)
	0-2 cm	15-21 cm					0-2 cm	15-21 cm			
0	19.0	14.0	5.0	0.0	-70.5	314	16.0	4.5	11.5	60.1	-41.0
10	18.5	12.5	6.0	3.7	-71.0	324	12.0	6.5	5.5	59.8	-42.0
20	19.0	10.5	8.5	7.0	-68.5	334	13.5	3.0	10.5	57.1	-40.0
40	15.0	8.5	6.5	7.4	-61.5	344	13.0	3.0	10.0	55.3	-40.0
50	15.0	8.8	6.2	16.7	-65.0	354	11.0	2.0	9.0	50.0	-36.0
60	15.5	9.0	6.5	21.9	-64.5	364	12.0	3.5	8.5	47.8	-41.0
70	13.0	7.0	6.0	22.7	-63.5	374	12.5	3.0	9.5	50.0	-40.0
80	14.0	6.9	7.1	25.4	-63.5	384	12.0	3.0	9.0	49.2	-41.5
90	11.5	6.5	5.0	27.6	-64.0	394	14.5	4.0	10.5	48.0	-41.0
100	14.0	6.3	7.7	30.9	-63.5	404	11.5	4.2	7.3	46.7	-45.0
110	11.5	6.0	5.5	32.1	-61.5	414	9.0	5.0	4.0	46.5	-44.0
120	15.0	6.0	9.0	37.3	-60.5	424	10.0	4.0	6.0	45.7	-49.5
125	13.0	6.0	7.0	38.2	-56.5	434	16.0	4.0	12.0	48.4	-49.5
130	12.0	6.0	6.0	39.1	-54.5	444	13.5	5.5	8.0	47.2	-47.5
140	15.0	5.0	10.0	40.3	-50.5	454	12.0	4.0	8.0	46.4	-46.0
150	11.0	4.2	6.8	40.5	-46.5	464	11.0	3.5	7.5	46.2	-45.5
160	12.5	3.7	8.8	42.8	-41.0	474	9.5	3.7	5.8	45.4	45.0
170	16.0	3.5	12.5	47.0	-41.0	484	8.5	3.5	5.0	45.1	-45.0
180	15.0	4.0	11.0	50.7	-40.0	494	11.0	4.0	7.0	50.9	-47.0
190	14.5	5.5	9.0	50.5	-36.0	504	12.0	4.0	8.0	52.1	48.0
200	17.0	4.4	12.6	52.2	-38.5	514	17.0	4.0	13.0	51.3	-47.0
210	19.0	5.0	14.0	50.0	-32.0	524	13.0	2.5	10.5	41.6	-32.0
220	16.0	5.0	11.0	49.7	-27.0	534	11.0	2.5	8.5	42.3	-32.0
230	14.0	4.2	9.8	46.1	-26.5	544	12.5	2.2	10.3	44.1	-32.0
240	12.0	3.3	8.7	46.2	-26.0	554	12.0	4.0	8.0	42.3	-26.0
250	17.0	6.0	11.0	50.9	-28.5	564	9.0	2.2	6.8	37.0	-35.0
260	15.0	4.0	11.0	53.6	-25.0	574	14.0	5.5	8.5	41.3	-33.0
270	12.0	1.5	10.5	49.4	-21.0	584	13.0	6.0	7.0	44.0	-37.5
280	12.5	2.0	10.5	47.6	-21.5	594	18.0	4.0	14.0	44.3	-39.0
290	15.0	3.0	12.0	57.3	-34.5	604	17.0	2.5	14.5	42.0	-38.0
300	12.5	3.5	9.0	60.1	-37.5	614	17.0	2.0	15.0	41.2	-38.0

Note: F.T. is the depth of the frost table relative to the surface at X = 0

7.3. Field data from North Point field site

Location: Slope 200 m south of Bore Hole 1 Air Temp.: 15.4°C
 Date: 28/7/90 Bearing of transect: 214°
 Time: 1:25 pm (magnetic declination of 38° assumed)

X (m)	Temperature (°C)		Zt (cm)	ΔT (°C)	Zsfc (m)	Frost Table (m)	SnowDepth (m)	SnowSic (m)
	0-7 cm	14-21 cm						
0.00	-	-	-	-	0.00	-	0.24	0.24
1.52	-	-	-	-	0.00	-	0.17	0.17
3.35	-	-	-	-	0.00	-	0.15	0.15
5.49	-	-	-	-	0.00	-	0.17	0.17
9.14	-	-	-	-	0.00	-	0.23	0.23
10.06	-	-	-	-	0.00	-	0.33	0.33
11.73	-	-	-	-	0.00	-	0.46	0.46
13.72	-	-	-	-	0.00	-	0.63	0.63
15.24	-	-	-	-	0.00	-	0.83	0.83
18.29	-	-	-	-	0.00	-	1.09	1.09
21.34	-	-	-	-	0.00	-	1.37	1.37
22.86	-	-	-	-	0.00	-	1.66	1.66
24.38	-	-	-	-	0.00	-	1.91	1.91
27.43	-	-	-	-	0.00	-	2.14	2.14
30.48	-	-	-	-	0.00	-	2.40	2.40
30.90	-	-	-	-	0.00	-	2.47	2.47
31.00	-	-	-	-	0.65	-	1.86	2.51
31.60	11.5	4.0	55.0	7.5	0.70	0.15	1.87	2.57
32.09	9.2	4.0	53.0	5.2	0.81	0.28	1.78	2.59
32.58	10.0	3.9	58.5	6.1	0.90	0.32	1.69	2.59
33.08	8.2	3.9	59.0	4.3	0.94	0.35	1.64	2.58
33.57	8.2	3.5	54.5	4.7	0.99	0.45	1.60	2.59
34.06	10.0	5.0	60.0	5.0	1.12	0.52	1.51	2.63
34.55	8.5	4.8	60.0	3.7	1.14	0.54	1.55	2.69
35.05	11.0	5.9	60.5	5.1	1.22	0.62	1.55	2.77
35.54	9.1	3.9	54.0	5.2	1.28	0.74	1.59	2.87
36.52	9.2	3.5	53.0	5.7	1.45	0.92	1.53	2.98
37.01	11.0	5.5	62.5	5.5	1.64	1.02	1.45	3.09
37.51	11.5	6.0	62.0	5.5	1.74	1.12	1.44	3.18
38.00	11.2	5.5	71.0	5.7	1.86	1.15	1.42	3.28
38.49	7.5	3.5	49.0	4.0	1.89	1.4	1.49	3.38
38.99	9.5	4.9	62.5	4.6	1.98	1.36	1.47	3.45
39.48	9.0	4.9	62.5	4.1	2.03	1.4	1.48	3.51
39.98	10.5	4.5	63.5	6.0	2.10	1.47	1.46	3.56
40.47	8.9	5.5	63.0	3.4	2.19	1.56	1.42	3.61
40.96	10.9	5.0	61.0	5.9	2.24	1.63	1.40	3.64
41.46	9.2	4.8	58.0	4.4	2.27	1.69	1.42	3.69
41.96	9.0	4.5	58.0	4.5	2.34	1.76	1.39	3.73
42.46	7.0	3.8	51.5	3.2	2.36	1.85	1.42	3.78
42.96	8.8	3.4	46.0	5.4	2.39	1.93	1.44	3.83

X (m)	Temperature (°C)		Zt (cm)	ΔT (°C)	Zsfc (m)	Frost Table (m)	SnowDepth (m)	SnowStc (m)
	0-7 cm	14-21 cm						
65.05	9.5	3.0	12.0	6.4	6.41	6.04	0.59	6.97
65.54	9.0	4.0	35.5	5.0	6.52	6.16	0.57	7.09
66.03	10.5	4.6	40.0	5.9	6.69	6.29	0.53	7.22
66.51	8.5	4.5	45.0	4.0	6.81	6.36	0.54	7.35
67.00	10.5	4.5	44.5	6.0	6.92	6.48	0.54	7.46
67.49	11.0	6.0	54.5	5.0	7.10	6.56	0.47	7.57
67.98	10.0	7.5	56.0	2.5	7.25	6.69	0.41	7.66
68.47	10.5	7.0	58.0	3.5	7.40	6.82	0.33	7.73
68.96	9.0	4.0	48.0	5.0	7.39	6.91	0.39	7.78
69.44	8.5	4.5	47.5	4.0	7.47	7	0.37	7.84
69.93	9.0	5.5	53.0	3.5	7.68	7.15	0.22	7.90
70.42	9.5	7.0	48.0	2.5	7.69	7.21	0.28	7.97

Location: Slope 200 m south of Bore Hole 1 Air Temp.: 12.7°C
Date: 29/7/90 Bearing of transect: 214°
Time: 10:10 am (magnetic declination of 38° assumed)

X (m)	Temperature (°C)		Zt (cm)	ΔT (°C)	Zsfc (m)	Frost Table (m)	SnowDepth (m)	SnowStc (m)
	0-7 cm	14-21 cm						
70.90	7.5	4.3	42.5	3.2	7.79	7.37	0.28	8.07
71.39	7.6	2.9	36.0	4.7	7.87	7.51	0.32	8.19
71.88	7.0	4.4	42.0	2.6	8.08	7.66	0.25	8.33
72.37	8.5	4.9	40.0	3.6	8.16	7.76	0.30	8.46
72.85	7.5	5.0	46.0	2.5	8.32	7.86	0.27	8.59
73.34	8.0	6.5	59.5	1.5	8.48	7.89	0.24	8.72
73.83	10.1	6.9	63.5	3.2	8.58	7.95	0.24	8.82
74.33	8.7	6.0	57.0	2.7	8.62	8.05	0.30	8.92
74.82	8.5	5.5	56.0	3.0	8.69	8.13	0.30	8.99
75.31	8.6	5.6	53.0	3.0	8.75	8.22	0.30	9.05
75.80	8.6	5.1	51.0	3.5	8.78	8.27	0.34	9.12
76.30	8.8	4.0	46.0	4.8	8.76	8.3	0.44	9.20
76.78	9.0	4.7	46.5	4.3	8.93	8.46	0.35	9.28
77.27	7.2	4.0	42.0	3.2	9.01	8.59	0.38	9.39
77.76	7.5	3.8	46.0	3.7	9.16	8.7	0.36	9.52
78.24	7.4	4.8	48.5	2.6	9.32	8.84	0.33	9.65
78.73	7.5	4.9	51.0	2.6	9.44	8.93	0.34	9.78
79.22	8.0	4.5	43.0	3.5	9.47	9.04	0.45	9.92
79.70	9.5	6.0	59.5	3.5	9.77	9.17	0.28	10.05
80.18	7.1	4.5	53.5	2.6	9.82	9.29	0.37	10.19
80.67	8.6	4.0	46.0	4.6	9.85	9.39	0.45	10.30
81.15	8.5	4.0	49.5	4.5	9.98	9.49	0.43	10.41
81.64	8.8	7.2	64.0	1.6	10.23	9.59	0.29	10.52
82.12	9.8	5.6	51.0	4.2	10.21	9.7	0.42	10.63

X (m)	Temperature (°C)		Z1 (cm)	ΔT (°C)	Zslc (m)	Frost Table (m)	SnowDepth (m)	SnowSlc (m)
	0-7 cm	14-21 cm						
43.45	9.9	5.5	51.0	4.4	2.48	1.97	1.41	3.89
43.95	10.3	4.0	56.0	6.3	2.59	2.03	1.37	3.96
44.44	11.2	3.7	49.0	7.5	2.69	2.2	1.34	4.03
44.93	7.5	4.5	56.0	3.0	2.81	2.25	1.30	4.11
45.42	9.0	3.8	52.0	5.2	2.86	2.34	1.34	4.20
45.91	13.0	4.5	56.0	8.5	3.00	2.44	1.28	4.28
46.41	8.0	3.8	47.0	4.2	2.99	2.52	1.37	4.36
46.90	8.5	4.2	50.0	4.3	3.13	2.63	1.30	4.43
47.39	10.2	4.0	53.0	6.2	3.25	2.72	1.26	4.51
47.88	10.5	4.5	47.5	6.0	3.32	2.84	1.27	4.59
48.37	11.0	3.7	45.0	7.3	3.41	2.96	1.26	4.67
48.86	13.1	7.0	45.5	6.1	3.51	3.05	1.23	4.74
49.35	11.8	4.5	44.0	7.3	3.55	3.11	1.29	4.84
49.84	8.5	4.2	50.0	4.3	3.69	3.19	1.23	4.92
50.33	8.0	3.6	45.0	4.4	3.72	3.27	1.29	5.01
50.82	11.2	4.0	45.0	7.2	3.83	3.38	1.28	5.11
51.31	11.5	4.0	44.5	7.5	3.93	3.49	1.25	5.18
51.80	11.0	5.0	53.0	6.0	4.08	3.55	1.17	5.25
52.29	12.5	5.0	50.0	7.5	4.12	3.62	1.20	5.32
52.79	9.5	4.7	57.0	4.8	4.26	3.69	1.12	5.38
53.28	12.0	4.0	45.5	8.0	4.24	3.79	1.19	5.43
53.77	9.5	4.5	44.5	5.0	4.29	3.85	1.18	5.47
54.27	9.2	5.0	56.5	4.2	4.49	3.93	1.01	5.50
54.76	12.0	5.2	57.5	6.8	4.58	4.01	0.95	5.53
55.26	11.3	6.0	56.5	5.3	4.63	4.06	0.93	5.56
55.75	10.5	3.8	49.0	6.7	4.66	4.17	0.95	5.61
56.24	12.2	3.9	46.0	8.3	4.74	4.28	0.93	5.67
56.72	8.5	4.5	50.5	4.0	4.87	4.37	0.84	5.71
57.21	9.5	3.5	48.0	6.0	4.95	4.47	0.79	5.74
57.70	13.0	3.0	43.5	10.0	5.06	4.63	0.73	5.79
58.19	10.5	4.4	49.5	6.1	5.23	4.74	0.61	5.84
58.67	6.5	2.5	39.0	4.0	5.18	4.79	0.72	5.90
59.17	6.5	1.5	9.7	5.2	5.26	4.91	0.75	5.97
59.66	7.2	1.1	9.7	5.3	5.38	4.98	0.77	6.05
60.15	9.5	3.2	33.5	6.3	5.45	5.12	0.70	6.15
60.64	11.8	3.0	35.0	8.8	5.53	5.18	0.71	6.24
61.14	9.5	5.0	48.5	4.5	5.73	5.25	0.59	6.32
61.63	16.0	5.0	48.0	11.0	5.86	5.38	0.56	6.42
62.12	14.0	4.8	44.5	9.2	5.95	5.51	0.57	6.52
62.61	6.0	2.0	33.0	4.0	5.96	5.63	0.64	6.60
63.09	8.5	2.5	33.0	6.0	6.06	5.73	0.59	6.65
63.58	8.0	3.0	31.5	5.0	6.16	5.85	0.56	6.72
64.07	8.0	4.0	42.0	4.0	6.34	5.92	0.44	6.78
64.56	9.5	3.8	39.0	5.7	6.36	5.97	0.51	6.87

X (m)	Temperature (°C)		Zt (cm)	ΔT (°C)	Zsfc (m)	Frost Table (m)	SnowDepth (m)	SnowSfc (m)
	0-7 cm	14-21 cm						
82.61	9.2	5.4	56.0	3.8	10.33	9.77	0.38	10.71
83.11	9.0	3.6	45.5	5.4	10.32	9.87	0.48	10.80
83.60	8.6	6.0	59.0	2.6	10.56	9.97	0.34	10.90
84.10	9.0	5.4	57.5	3.6	10.61	10.04	0.37	10.98
84.59	8.0	5.2	58.5	2.8	10.71	10.13	0.36	11.07
85.09	9.0	5.1	54.0	3.9	10.77	10.23	0.39	11.16
85.57	9.2	6.0	60.0	3.2	10.94	10.34	0.30	11.24
86.06	8.5	6.2	63.0	2.3	11.06	10.43	0.25	11.31
86.54	9.0	4.7	54.0	4.3	11.07	10.53	0.32	11.39
87.02	8.2	5.6	55.0	2.6	11.21	10.66	0.26	11.47
87.51	8.0	5.0	44.5	3.0	11.22	10.78	0.34	11.56
87.99	7.4	4.2	48.0	3.2	11.35	10.87	0.31	11.66
88.49	9.7	5.4	55.0	4.3	11.51	10.96	0.24	11.75
88.98	10.1	6.0	63.0	4.1	11.68	11.05	0.17	11.85
89.47	8.0	4.5	53.0	3.5	11.70	11.17	0.25	11.95
89.96	10.0	6.0	53.5	4.0	11.83	11.3	0.22	12.05
90.46	10.0	6.5	64.0	3.5	11.97	11.33	0.19	12.16
90.95	7.7	4.0	51.0	3.7	11.90	11.39	0.36	12.26
91.44	9.9	5.2	57.0	4.7	12.04	11.47	0.32	12.36
92.44	8.6	5.0	59.5	3.6	12.21	11.62	0.24	12.45
93.43	10.6	5.1	58.0	5.5	12.35	11.77	0.18	12.53
93.92	8.2	5.1	55.0	3.1	12.42	11.87	0.18	12.60
94.42	14.0	6.0	54.5	8.0	12.47	11.93	0.20	12.67
95.42	10.1	4.6	48.0	5.5	12.47	11.99	0.27	12.74
96.41	9.8	4.0	47.5	5.8	12.49	12.02	0.29	12.78
96.91	10.0	5.0	58.0	5.0	12.55	11.97	0.26	12.81
97.91	10.0	5.1	58.5	4.9	12.67	12.08	0.20	12.87
98.91	8.3	7.1	48.0	1.2	12.67	12.19	0.26	12.93
99.91	10.0	4.5	52.0	5.5	12.73	12.21	0.29	13.02
100.90	10.5	5.5	55.0	5.0	12.87	12.32	0.23	13.10
101.89	10.1	4.9	57.5	5.2	12.96	12.39	0.22	13.18
102.89	8.5	4.7	64.5	3.8	13.11	12.47	0.16	13.27
103.89	12.0	3.3	50.0	8.7	13.08	12.58	0.24	13.32
104.89	11.5	5.0	53.0	6.5	13.22	12.69	0.14	13.36
105.89	9.0	3.8	44.0	5.2	13.13	12.69	0.28	13.41
106.89	11.0	5.2	56.5	5.8	13.27	12.71	0.16	13.43
107.89	9.0	5.1	55.0	3.9	13.27	12.72	0.18	13.45
108.88	10.2	5.5	44.0	4.7	13.14	12.7	0.33	13.47
109.88	10.5	5.1	57.0	5.4	13.31	12.74	0.18	13.49
110.88	11.2	6.0	62.5	5.2	13.44	12.82	0.09	13.53
111.88	9.2	4.2	46.5	5.0	13.31	12.85	0.26	13.57

END

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