N-FACTORS AND THE RELATIONS BETWEEN AIR AND SURFACE
TEMPERATURE IN DISCONTINUOUS PERMAFROST NEAR MAYO, YUKON
TERRITORY

by

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A thesis submitted to
the Faculty of Graduate Studies
in partial fulfillment of
the requirements of the degree of

Master of Science

Carleton University
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“N-FACTORS AND THE RELATIONS BETWEEN
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Thesis Supervisor

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ABSTRACT

Air and surface temperatures were measured for one year at five sites with different permafrost conditions near Mayo, Yukon Territory, to examine the behaviour of freezing and thawing n-factors and the relations between air and surface temperatures. During the freezing season, surface temperatures were lower where the snow cover was thin. Where snow covers were similar, surface temperatures were lower where permafrost was present. The freezing n-factor is controlled primarily by snow, but also by subsurface thermal conditions. During the thawing season, surface temperatures were higher where the near-surface thermal diffusivity allowed for rapid descent of the frost table. The thawing n-factor is primarily controlled by near-surface thermal diffusivity. The freezing n-factor is representative of the relation between air and surface temperature only during the mid season, while the thawing n-factor is representative of mid-season conditions when the mid season is longer than the early season.

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CHAPTER ONE

OVERVIEW AND OBJECTIVES

1.1 Introduction

This thesis examines air and ground surface temperatures in the discontinuous permafrost of central Yukon Territory. Permafrost is ground that remains at or below 0°C for at least two consecutive years (ACGR, 1988). Permafrost is a primary component of northern environmental systems because it affects vegetation growth (Zoltai, 1975), geochemical cycling (Kokelj *et al.*, 2002), terrain stability (Lewkowicz, 1990) and the hydrologic cycle (Woo and McCann, 1994).

The permafrost region in Canada has been broadly subdivided into the continuous and discontinuous permafrost zones (Figure 1.1) (Heginbottom *et al.*, 1995). The climate in the discontinuous permafrost zone is insufficient for permafrost to be both spatially and temporally continuous (Brown, 1969). Rather, the existence of permafrost there also depends on site characteristics, i.e. microclimatic factors. Permafrost conditions in the region are sensitive to changes in climate or site characteristics because the ground temperature is close to 0 °C. Understanding of the ground thermal regime in the discontinuous permafrost zone is necessary to reduce the environmental impact of economic development and to monitor the affects of climate change.

1.2 Permafrost modeling

Modeling permafrost distribution involves determining the mean annual ground temperature. Using numerical models of heat conduction, mean annual ground temperature can be determined from the mean annual surface temperature (Smith and Riseborough, 1996; Klene *et al.*, 2001). Direct mapping of the ground thermal regime

Figure 1.1 Distribution of permafrost in Canada (after Heginbottom *et al.* 1995, M.M. Côté, 2002, Fig. 2.2).

throughout the discontinuous permafrost zone requires mean annual surface or ground temperature measurements from numerous locations. However, surface and ground temperature records are only available from a few locations and therefore do not form a practical basis for regional mapping. Air temperature records, on the other hand, are available for many communities and other sites throughout the discontinuous permafrost zone, and are potentially useful for such mapping. However, to use air temperature in heat conduction models, air temperature must first be related to ground surface temperature, because heat energy transferred between the surface and the air is primarily by convection rather than conduction.

1.3 The n-factor

Air and ground surface temperatures are related through a series of energy fluxes that compose the surface energy balance. To relate air and ground surface temperature requires quantification of the energy balance, which is time and instrument intensive. Instead the *n-factor*, a ratio of the ground surface freezing/thawing index to the air freezing/thawing index has been developed to summarize the surface energy balance. The n-factor is used to estimate ground surface temperature from air temperature measurements (Lunardini, 1978). The n-factor is generated for the winter (freezing n-factor, n_f), and for the summer (thawing n-factor, n_t) usually using freezing or thawing degree-days for the surface and air (Lunardini, 1978). The n-factor has been used in numerical models of the ground thermal regime to calculate surface temperatures from air temperature (Jorgenson and Krieg, 1988; Smith and Riseborough, 1996; Wright *et al.*, 2000; Klene *et al.*, 2001).

Understanding the behaviour of n-factors is critical for accurate modeling of the ground thermal regime. However, several aspects of n-factor behaviour have not been examined in detail (Taylor, 1995; Klene *et al.*, 2001). First, while the n-factor is controlled by site characteristics, the specific characteristics influencing the n-factor have not been identified. Understanding such relations would improve the assignment of n-factors to surfaces of interest. Second, the temporal variability of the n-factor within the freezing and thawing seasons has not been examined. While the n-factor is defined on a seasonal basis, the behaviour of the n-factor throughout the seasons is important because the relation between air and surface temperature changes over the freezing and thawing seasons, due to changes in vegetation canopy during summer, and snow cover during winter.

1.4 Research Objectives

The purpose of this research was to determine how spatial and temporal variations in the relation between air and surface temperature are affected by site characteristics. Air and surface temperatures were measured for one year at five sites in the discontinuous permafrost zone near Mayo, Yukon Territory. The sites exhibited different surface and sub-surface characteristics, but are all located within a small area (3 km²). The relations between air and surface temperature and the physical characteristics of the sites were examined. The research objectives were:

1. To describe the differences in air and surface temperature between the study sites.

2. To produce thawing and freezing n-factors for five study sites in discontinuous permafrost near Mayo, YT.
3. To determine the physical characteristics that control the freezing and thawing n-factors at the study sites.
4. To describe the relations between air and surface temperature over the freezing and thawing seasons at the five study sites.
5. To determine how representative the n-factor is of the ratio of surface to air temperature throughout the thawing and freezing seasons at the five study sites.

1.5 Structure of the Thesis

This thesis is composed of seven chapters. The next chapter discusses how n-factors summarize the relation between surface and air temperatures. Chapter Three describes the study area, the field methods used, and how n-factors were derived. Chapter Four describes the physical characteristics of the study sites. Chapter Five examines the air and surface temperatures at the study sites and presents freezing and thawing n-factors for the study sites. Chapter Six examines the relation between air and surface temperature throughout the freezing and thawing seasons at the study sites. Chapter Seven summarizes the results of this thesis, presents conclusions, and suggests opportunities for future research.

CHAPTER TWO

N-FACTORS AND THE RELATION BETWEEN AIR AND SURFACE TEMPERATURE

2.1 Introduction

An examination of n-factors in the discontinuous permafrost zone requires an understanding of (1) how air and surface temperatures are related through the surface energy balance, (2) how the surface energy balance is affected by site characteristics, (3) how n-factors are used to summarize the surface energy balance, and (4) the current knowledge of their behaviour. This chapter reviews the relations between air and surface temperature in the discontinuous permafrost zone and the calculation, application and behaviour of n-factors.

2.2 Surface energy exchanges

The surface energy balance describes how radiant energy from the sun is partitioned at the ground surface (Oke, 1978). The radiation balance quantifies fluxes of radiant energy, and determines the net radiation (Q^*) at the surface. The surface energy balance is driven by Q^* and is comprised of fluxes of thermal energy.

2.2.1 The radiation balance

Emitted radiation is a function of a body's temperature and emissivity. Emissivity is the ability to absorb and emit energy. The Earth's radiation balance involves short-wave radiation (K), and long-wave radiation (L). Short-wave radiation is emitted by the sun, while long-wave radiation is emitted by the earth-atmosphere system

(Oke, 1978). The radiation balance describes how radiant energy incident on the surface is absorbed, reflected and reemitted, and is expressed as:

$$Q^* = (K\downarrow - K\uparrow) + (L\downarrow - L\uparrow) \quad [2.1]$$

where $K\downarrow$ is incoming short-wave radiation, $K\uparrow$ is outgoing short-wave radiation, $L\downarrow$ is incoming long-wave radiation, and $L\uparrow$ is outgoing long-wave radiation. Q^* is the resulting surplus or deficit of radiation, and is spatially and temporally variable because the magnitude of the radiative fluxes is controlled by variable surface and atmospheric characteristics.

2.2.1.1 Controls on incoming short-wave radiation ($K\downarrow$)

Short-wave radiation is emitted from the sun. The magnitude of $K\downarrow$ is highly dependent on the position of the sun in the sky (Oke, 1978). Therefore $K\downarrow$ varies throughout the day, being highest at noon, and low at sunrise and sunset. At high northern latitudes ($>60^\circ$ N), the path that the sun takes across the sky varies greatly throughout the year (Sellers, 1965). During winter, the sun's path across the sky is short which results in low $K\downarrow$. Conversely, during summer, the sun's path across the sky is long which results in high $K\downarrow$.

Increases in cloud cover and atmospheric constituents result in a decrease in $K\downarrow$ because more of the short-wave radiation entering the atmosphere is absorbed, scattered, or reflected back out of the atmosphere (Oke, 1978). Objects can also reduce $K\downarrow$ by blocking incoming radiation. In natural environments vegetation, mountains and other landforms often block $K\downarrow$ (Oke, 1978).

2.2.1.2 Controls on incoming long-wave radiation ($L\downarrow$)

$L\downarrow$ is emitted either by the atmosphere or by local surface objects. Long-wave radiation emitted by the atmosphere is controlled by the emissivity, temperature, and concentration of atmospheric constituents, especially water vapour, carbon dioxide, and ozone (Oke, 1978). Other objects, such as vegetation, can also contribute $L\downarrow$ to the surface. Their contribution is controlled by the emissivity, temperature, and size of the object, as well as its proximity to the receiving surface.

2.2.1.3 Controls on outgoing short-wave radiation ($K\uparrow$)

$K\uparrow$ is short-wave radiation reflected off a surface. The magnitude of $K\uparrow$ is dependent upon the magnitude of $K\downarrow$ and the surface reflectivity, or albedo. The albedo is simply the ratio of $K\downarrow$ to $K\uparrow$, and thus ranges from 0 to 1 (Oke, 1978). As the albedo of a surface increases, $K\uparrow$ increases. While natural surface materials have a wide range of albedos, light-coloured materials, such as snow tend to have high albedos (Table 2.1). In contrast, dark wet soils and coniferous forests have low albedos (Table 2.1).

2.2.1.4 Controls on outgoing long-wave radiation ($L\uparrow$)

$L\uparrow$ is long-wave radiation emitted by a surface. The magnitude of $L\uparrow$ is directly related to the emissivity and temperature of the surface through the Stefan-Boltzmann Law:

$$L\uparrow = \epsilon \sigma T^4 \quad [2.2]$$

where ϵ is the surface emissivity, σ is the Stefan-Boltzmann constant, and T is the surface temperature. Since natural surface materials have a narrow range of emissivities

Table 2.1 Albedo and emissivity of natural surface materials (after Oke, 1978, Table 1.1, p. 12).

Surface	Remarks	Albedo	Emissivity
Soils	Dark, wet to Light, dry	0.05 to 0.40	0.98 to 0.90
Grass	Long (1 m) to Short (0.02 m)	0.16 to 0.26	0.90 to 0.95
Forests	Deciduous	0.15 - 0.20	0.97 - 0.98
	Coniferous	0.05 - 0.15	0.97 - 0.99
Water	Small zenith angle	0.03 - 0.10	0.92 - 0.97
	Large zenith angle	0.10 - 1.00	0.92 - 0.97
Snow	Old to Fresh	0.40 to 0.95	0.82 to 0.99

(Table 2.1), the magnitude of $L\uparrow$ is predominantly controlled by surface temperature (Oke, 1978).

2.2.1.5 Net radiation (Q^*)

Q^* is positive (surplus) when incoming radiation is greater than outgoing radiation, and negative (deficit) when outgoing radiation is greater than incoming radiation. The value of Q^* determines the magnitude and direction of the thermal energy fluxes that make up the surface energy balance.

2.2.2 The surface energy balance

The surface energy balance describes fluxes of energy towards or away from the surface:

$$Q^* = Q_E + Q_H + Q_G \quad [2.3]$$

where Q_E is the latent heat flux, Q_H is the sensible heat flux, and Q_G is the ground heat flux (Oke, 1978). Q_E is the heat released or captured during phase changes of water. The latent heat of vaporization involves transfer of both heat and water between the surface and overlying air, while the latent heat of fusion only involves the transfer of heat. Sensible heat is a measure of the heat content of a body, in this case the surface. Q_H is transferred by convection to or from the overlying air. Q_G is heat conducted at the surface to or from the underlying soil.

Q^* controls the direction of the thermal fluxes, either away from the surface when Q^* is positive, or towards the surface when Q^* is negative (Oke, 1978). At all times the

sum of Q_E , Q_H , and Q_G must equal Q^* , but the relative partitioning of Q_E , Q_H , and Q_G is dependent upon surface characteristics.

2.2.2.1 Controls on the latent heat flux (Q_E)

The magnitude of Q_E is controlled primarily by the availability of surface water (Oke, 1978). When the surface is dry, Q_E is negligible. The latent heat of vapourization is also controlled by the vapour concentration gradient between the surface and the overlying air, which is dependent on atmospheric turbulence. As the vapour gradient increases, Q_E increases.

At the surface, the latent heat of fusion is significant in cold climates for only a short period in the spring and fall as the surface water melts and freezes. Latent heat released or absorbed at depth is considered part of Q_G .

2.2.2.2 Controls on the sensible heat flux (Q_H)

Q_H is controlled by the temperature gradient between the surface and the overlying air; as the temperature gradient increases, Q_H increases. The temperature gradient is dependent on atmospheric turbulence, which transports heated air away from the surface (Oke, 1978).

The partitioning of Q^* between Q_H and Q_E is described by the *Bowen ratio* (β) which is expressed as:

$$\beta = Q_H / Q_E \quad [2.4]$$

Thus, high values of β indicate that $Q_H > Q_E$ and *vice versa*. Deserts and other dry surfaces have high values of β , while β values are low for wetlands

2.2.2.3 Controls on the ground heat flux (Q_G)

The partitioning of heat transfer between Q_G and Q_H is controlled by the surface and the atmospheric thermal admittance. Thermal admittance describes the ease with which a body releases or takes up heat. Thermal admittance is controlled by the heat capacity of the air or soil, and the ability of the air or soil to transfer heat.

The magnitude of Q_G is controlled by the temperature gradient between the surface and the sub-surface, and the thermal conductivity of the soil (Oke, 1978). Thermal conductivity is the ability of a soil to conduct heat. The conductivity of a soil is determined by the thermal conductivity of the soil constituents (Table 2.2) (Oke, 1978). Increasing the soil moisture increases the thermal conductivity of a soil until saturation, when the thermal conductivity will remain constant.

The rate of temperature change in soil is described by the thermal diffusivity of the soil, which is the ratio of the thermal conductivity to the heat capacity (Oke, 1978). The heat capacity is the energy required to raise a unit volume by 1°C . The heat capacity of a soil is highly dependent on moisture and organic content (Table 2.2). As soil moisture increases, the thermal diffusivity rises due to the increase in thermal conductivity; however, as the soil approaches saturation, the thermal diffusivity decreases because the thermal conductivity levels off, but the heat capacity continues to rise. The thermal diffusivity is limited at low soil moisture contents by the low thermal conductivity, and at high moisture contents by the high heat capacity (Oke, 1978). When soil water freezes, the thermal diffusivity increases because the thermal conductivity of ice is higher than that of water, and the heat capacity is lower (Table 2.2).

Table 2.2 Thermal properties of soil constituents and snow (after Oke, 1978, Table 2.1, p.44 ; and Williams and Smith, 1989, Table 4.1, p.90).

	Heat Capacity (J kg ⁻¹ K ⁻¹)	Thermal Conductivity (Wm ⁻¹ K ⁻¹)	Thermal Diffusivity (x10 ⁻⁶ m ² s ⁻¹)	Thermal Addmittance (J m ² s ^{-1/2} K ⁻¹)
Soil Constituents				
Quartz	800	8.80	4.14	-
Clay minerals	900	2.92	1.22	-
Organic matter	1920	0.25	0.10	-
Water (0°C)	4180	0.56	0.13	1545
Ice (0°C)	2100	2.24	1.16	2080
Air (still)	1010	0.025	20.63	5
Snow				
Fresh snow	2090	0.08	0.10	130
Old snow	2090	0.42	0.40	595

2.2.3 Relations between air and surface temperature

Air temperature is controlled by climate, which is both dynamic and physical (Hare, 1997). The dynamic component of climate is that which is imported or advected from other regions. The physical component is local or the product of the surface energy balance.

Both air and surface temperature control the thermal gradient for Q_H . Therefore, air and surface temperature are interdependent, and positively correlated. The strength of this correlation is dependent upon the height above the surface that the air temperature is measured (Oke, 1978). As the height increases, the surface area that influences the air temperature increases, which results in a decrease in the strength of the correlation between air and surface temperature. The correlation weakens if the surface characteristics are spatially variable. In addition, as the distance from the surface increases, the influence of the advected component of air temperature increases (Hare, 1997). This component is not reflective of the local surface conditions, but rather of surface conditions some distance away. The advected component also increases if the wind speed increases (Oke, 1978). Thus, as wind speed increases, the correlation between air and surface temperature weakens.

2.3 The surface energy balance in the boreal forest

The boreal forest is a northern forest biome, which covers 40% of Canada in a broad NW- SE band (Ritchie, 1993) (Figure 2.1). The northern boreal forest is dominated by coniferous tree species but deciduous trees are also common. Beneath the

Figure 2.1 Extent of boreal forest in Canada (Hare and Ritchie, 1972).

forest canopy, shrubs and herbaceous plants are present, and the ground surface is often covered with mosses and lichens (Larsen, 1980).

Due to its high latitude, winter lasts 6-8 months and has few daylight hours and cold conditions, while summer lasts 4-6 months and has long days and warm conditions (Larsen, 1980). Approximately one third of the annual precipitation in the boreal forest arrives as snow (Larsen, 1980). Snow depth is highly variable at a local scale due to interception by vegetation, especially coniferous trees, and redistribution by wind (Smith, 1975; Pomeroy and Goodison, 1997).

The northern limit of discontinuous permafrost roughly coincides with the northern border of the boreal forest (Brown, 1967). Throughout the boreal forest the near surface ground annually freezes and thaws, but where permafrost is present, only the layer above the permafrost thaws. This layer is called the *active layer*. Where the climate is sufficiently cold the development of permafrost is promoted by the presence of spruce trees, mosses, and moist conditions (Brown and Péwé, 1973).

2.3.1 Net radiation in the boreal forest

The magnitude of Q^* in the boreal forest varies spatially with surface characteristics. Variations in the magnitude of Q^* in the boreal forest during summer are primarily the result of differences in vegetative shading. Smith (1975) measured only 23% of the net radiation for an open site under a closed canopy spruce forest, compared to 85% received in a stand of 1.5 m high willows. Therefore, coniferous trees are more effective at blocking incoming radiation than deciduous trees. Variations in understory surface albedo also contribute to the variability of Q^* . The albedos are approximately

0.22 for lichen (Petzold and Rencz, 1975), and range from 0.20 to 0.25 for mosses (Roberts, 2002), and from 0.05 for dark soils to 0.40 for light soils (Oke, 1978).

During winter, Q^* is negative due to the short day length and the low intensity of $K\downarrow$ (Sellers, 1965). The magnitude of Q^* during winter remains spatially variable, partly due to surface albedo, which increases to approximately 0.8 on snow-covered surfaces. (Larsen, 1980). During winter, the effect of shading is less important because of the low levels of Q^* .

2.3.2 The latent heat flux (Q_E) in the boreal forest

During winter, Q_E in the boreal forest is negligible because most surface water is frozen. In early spring when the snow begins to melt, the magnitude of Q_E begins to rise and continues to increase until mid summer. Q_E reaches its maximum in July and August when Q^* is high and surface water is available from summer rainstorms. Throughout late summer and fall Q_E steadily declines due to the decrease in Q^* , and becomes negligible again when surface water freezes (Betts *et al.*, 2001).

The magnitude of Q_E varies throughout the boreal forest. Q_E is high over open wetlands with standing surface water. On vegetated surfaces Q_E is not only dependent on water availability, and atmospheric turbulence, but also on stomatal activity, which is a function of the species and condition of plants (Oke, 1987; Roulet *et al.*, 1997). Over moss-covered surfaces, Q_E is high in spring and after rainstorms, since surface water is readily available at these times. However, once the surface water has evaporated, Q_E often declines (Roulet *et al.*, 1997).

2.3.3 The sensible heat flux (Q_H) in the boreal forest

Variations in Q_H are best described with the Bowen ratio (β). During winter, β is high because Q_E is negligible, but begins to decline in April when snowmelt begins. In early summer β steadily declines as Q_E increases, but becomes stable in late summer when Q_E declines. As fall progresses, β increases again (Betts *et al.*, 2001). On a shorter time scale, β values decline immediately after rain events while surface water is evaporating (Oke, 1987). Typically, β for wetlands range from 0.40 to 0.85 (Roulet *et al.*, 1997) and are well above unity for dry forests (McCaughey *et al.*, 1997).

During winter, the low thermal conductivity of snow inhibits the transfer of heat from the ground surface to the air (Williams and Smith, 1989). This trapping of heat causes the surface temperature to remain warm despite cold air temperatures. The extent to which snow inhibits heat transfer increases with snow depth but decreases with snow density (Goodrich, 1982). The timing of the snow cover is also important to consider (Goodrich, 1982). During freeze-up, if the surface and underlying soil freeze before the surface is snow covered, the snow will preserve low surface temperatures once it arrives. Conversely, if the snow arrives before the surface is frozen, the snow cover will preserve high surface temperatures. In spring the surface cannot thaw until the overlying snow has melted, thus late-lying snow delays surface warming.

2.3.4 The ground heat flux in the boreal forest

The ground heat flux in the boreal forest is greatly affected by freezing and thawing of soil water. During the spring and summer, Q_G is spent warming the ground, and melting soil water. During the fall and winter, Q_G is negative and results in ground

cooling, and freezing of soil water. The rate of thawing or freezing decreases exponentially with time as the temperature gradient declines, however, the rate is dependent on soil water content. When soil water is freezing or thawing, ground temperature change is minimized due to exchanges of latent heat. When soil water freezes, latent heat is released and must be dissipated before ground cooling can continue. Likewise, the thawing of soil water requires heat, which therefore cannot be used to warm the ground (Williams and Smith, 1989). This period when the ground temperature does not change due to freezing or thawing is called the *zero-curtain* period (Williams and Smith, 1989).

The relation between thermal conductivity and temperature is dependent on the unfrozen water content of a soil (Williams and Smith, 1989). Unfrozen water is present in all freezing soils to varying degrees and is caused by the effects of solutes and adsorption that lower the freezing point. Thus, soil water freezes at a range of temperatures, especially in fine-grained soils that have a high amount of adsorbed solutes. In a coarse-grained soil, the unfrozen water content is very low because the majority of the soil water is not affected by solutes or adsorption and thus has a high freezing point. In a clay soil, the unfrozen water content changes gradually with decreasing temperature, therefore the thermal conductivity also changes gradually with decreasing temperature.

Peat greatly influences Q_G . Because of the high porosity of peat, its thermal conductivity is highly dependent on moisture conditions, which vary throughout the year. In summer, when soil moisture is low, the conductivity of peat is low, thus Q_G is reduced. In winter, the soil moisture is high due to rainfall in autumn, and the soil water is frozen. Since ice has a high conductivity, the conductivity of peat in winter is high. This

enhances Q_G . Thus peat has a strong but contrasting effect on Q_G in summer and winter (Williams and Smith, 1989). Non-conductive heat transfer is also considerable in peat due its high porosity. Nelson *et al.* (1985) propose that observations of simultaneous warming and cooling in peat at 10 and 20 cm respectively resulted from evaporation taking place at depth, and condensation nearer the surface.

Where permafrost is present, only the active layer rises above 0°C in summer. In summer the thermal gradient across the active layer is enhanced by the presence of permafrost as the temperature at the bottom of the thawed layer remains at 0°C (Rouse *et al.*, 1997; Klene, 2000). Thus permafrost can increase heat flow into the ground. Where permafrost is absent, Q_G is generally between 5 and 9% of Q^* and between 16 and 18% where permafrost is present (Rouse *et al.*, 1997).

Surface temperature can be predicted from air temperature if the surface energy balance is quantified. However, this approach is impractical. Only a small number of sites could be monitored due to the number of instruments needed to measure the radiative and thermal fluxes, and the extensive data processing required. The n-factor, an empirical ratio, is an alternative method of relating air to surface temperature.

2.4 The n-factor

Surface temperature can be estimated from air temperature using the n-factor, a ratio of the surface freezing/thawing index to the air freezing/thawing index. The n-factor summarizes the surface energy balance, and is attractive to those modeling the ground thermal regime because of its simplicity. Air and surface temperatures can be measured in several locations using inexpensive miniature data loggers, and n-factors can be readily

developed to represent numerous environments. As a result, the n-factor has become a common parameter in heat conduction models that attempt to estimate the ground thermal regime over large areas (Henry and Smith, 2000).

2.4.1 Deriving n-factors

Calculation of n-factors requires daily measurements of air and surface temperature from the same location. However, when n-factors are applied in numerical models, air temperatures measured at standard meteorological sites (usually at airports) are used to predict the temperature of nearby natural surfaces. If air temperatures from meteorological sites are used to predict the surface temperature of natural surfaces, errors may occur where the air temperature at a meteorological site is significantly different from that over a natural surface, i.e., a forest (Burn, 1998). These differences may be caused by differences in the surface energy balances and/or elevation.

Unlike air temperature, measurement of surface temperature is not standardized because it is difficult to define where the surface is. As a result, numerous methods have been used. Taylor (1995) and Klene *et al.* (2001) measured surface temperature by inserting a thermistor under the surface organic mat. Burn (1998) computed n-factors using soil temperature measured at a depth of 20 cm. Since the purpose of n-factors is to determine the boundary temperature for conductive heat transfer models, for n-factors, surface temperature should be measured at the interface between convective and conductive heat transfer. The interface between convective and conductive heat transfer usually coincides with the organic/mineral interface. However, during the winter the thermal properties of peat change considerably, therefore the position of the surface must

also change. Strictly speaking, the location where the surface temperature should be measured to derive the n-factor is, as yet, undefined.

The n-factor is computed on a seasonal basis: the freezing n-factor (n_f) is calculated when temperatures are below 0°C, and the thawing n-factor (n_t) is calculated when temperatures are above 0°C. The freezing/thawing indices from which the n-factors are calculated can be determined through integration of temperature over the freezing and thawing seasons. The n-factors then are:

$$n = \frac{\int_0^{\theta_s} (T_f - T_s) dt}{\int_0^{\theta_a} (T_f - T_a) dt} \quad [2.5]$$

where θ_s and θ_a are the lengths of the freezing or thawing seasons (in time t), T_s and T_a are the ground surface and air temperatures respectively, and T_f is the freezing temperature (0° C) (Lunardini, 1978).

Integration of temperature over the freezing or thawing season can be accomplished using freezing or thawing degree-days. Degree-days are the summation of temperature values over either the freezing season (FDD), or the thawing season (TDD) (Thompson, 1966). The n-factor is the ratio of ground surface degree-days to air degree-days for either the freezing season,

$$n_f = \frac{FDD_s}{FDD_a} \quad [2.6]$$

or the thawing season,

$$n_t = \frac{TDD_s}{TDD_a} \quad [2.7]$$

The degree-day method of calculating n-factors is widely used because it is easily understood and applied (Jorgenson and Krieg, 1988; Taylor, 1995; Klene, 2001).

Another method for determining n-factors involves finding the slope of the principal axis for the relation between air and surface temperature for each season (Burn, 1998). While this method is only valid if the intercept is zero, it allows for the calculation of the coefficient of determination (r^2), which gives an indication of the strength of the relation and thereby indicates the day-to-day variation of the n-factor. At a field site in Takhini valley, YT, thawing n-factors derived from the slope of the principal axis were similar to those derived using thawing degree-days, and the r^2 values were relatively high (Burn, 1998). Freezing n-factors derived from the slope of the principal axis were lower than those derived using freezing degree-days, and the r^2 values were low (Burn, 1998). Thus the principal axis is more reliable for determining thawing n-factors than freezing n-factors.

Regardless of the method used for deriving n-factors, the beginning and end of the freezing and thawing seasons must be defined. In equation 2.4 the air and surface temperature data sets are temporally independent. To improve the physical basis of the n-factor, the beginning and end of the freezing and thawing season should be the same for both air and surface temperature. However, definition of the n-factor season is not straightforward. Changes in surface temperature often lag changes in air temperature due to the higher heat capacity of the surface. In addition, season definition is difficult because of the variability about the freezing point during the spring and fall, especially in air temperature. Some published n-factors have been calculated using surface temperature to define the thawing and freezing seasons because of the lower variability of surface temperature (Taylor, 1995; Klene, 2001). The n-factor seasons have not been defined in other reports.

2.4.2 The application of n-factors

Originally n-factors were designed to predict thaw depth on artificial surfaces for engineering purposes (Carlson, 1952; Lunardini, 1978; Andersland and Ladanyi 1994). Jorgenson and Kreig (1988) first applied n-factors to natural surfaces to predict permafrost distribution within a small watershed. Their numerical model used n-factors, air degree-days and thermal conductivities for the freezing and thawing seasons to estimate the amount of heat entering the ground in summer versus the amount leaving in winter. Since then, the use of n-factors to predict ground thermal conditions in permafrost environments has expanded. For instance, Jorgenson and Kreig's model has been developed further to estimate permafrost thickness (Wright *et al.*, 2000).

Klene *et al.* (2001) incorporated thawing n-factors into a Stefan model to predict active-layer thickness and found an increase in the accuracy of the model. The n-factor is also used in a model that predicts the temperature at the top of permafrost (TTOP) from air temperature and subsurface thermal conditions (Smith and Riseborough, 1996). The TTOP model has been used to map ground temperatures (Henry and Smith, 2001), and to determine the relative influence of surface and subsurface conditions on ground temperatures (Riseborough and Smith, 1998; Smith and Riseborough, 2002). The integrity of these models is dependent upon the accurate assignment of n-factors to areas of interest.

2.4.3 The behaviour of n-factors

Since n-factors summarize the surface energy balance, they vary with site characteristics. Thus n-factors must be accompanied by detailed descriptions of the site

conditions that they represent so that they can be applied accurately to other areas of interest. Presently, site descriptions of published n-factors are not standardized, with variation in the site characteristics reported, and in level of detail included in the descriptions (Table 2.3). For example, while most descriptions include soil conditions, Jorgenson and Kreig (1988) reported soil moisture, Taylor (1995) reported parent material and occasionally geomorphic setting, while Klene *et al.* (2001) reported the soil classification. This variability reduces the ability to assign n-factors to other sites effectively, because it is unknown which site characteristics should be considered. An understanding of how site characteristics control the behaviour of n-factors would enable development of a standard method for describing the site that an n-factor represents and thereby would allow n-factors to be assigned more effectively to new areas.

2.4.3.1 Site characteristics controlling the n-factor

A number of studies have identified site characteristics that control the n-factor. Klene *et al.* (2001) compared n-factors from many different land-cover classes in the continuous permafrost zone of northern Alaska, and found that the thawing n-factor was low where soil moisture was high because of high Q_E . In contrast, the thawing n-factor was higher where there was standing surface water due to the low albedo of water, which increased Q^* .

Vegetation has also been identified as possibly controlling the thawing n-factor by blocking incoming radiation. Taylor (1995) found that in the Mackenzie Valley, the thawing n-factor was lower in forests than in open areas, and shading by shrubs has also been associated with lower thawing n-factors in tundra landscapes (Klene *et al.*, 2001).

Table 2.3 Published n-factors and associated site descriptions.

Site description	Soil Conditions	Nf	Nt	Reference
Spruce trees, brush and moss	peat	0.29	0.37	Carlson, 1952
Cleared of trees and brush, moss remains	peat	0.25	0.73	Carlson, 1952
Cleared of trees, brush, moss, and peat	silt loam	0.33	1.22	Carlson, 1952
Closed aspen forest	moisture= 15% wt.	0.3	1	Jorgenson and Kreig, 1988
Closed birch forest	moisture= 20% wt.	0.35	0.9	Jorgenson and Kreig, 1988
Mixed birch-spruce forest	moisture= 20% wt.	0.35	0.85	Jorgenson and Kreig, 1988
Closed white spruce forest	moisture= 50% wt.	0.35	0.8	Jorgenson and Kreig, 1988
Open black spruce forest	moisture= 50% wt.	0.3	0.6	Jorgenson and Kreig, 1988
Closed black spruce forest	moisture= 45% wt.	0.3	0.5	Jorgenson and Kreig, 1988
Low shrub scrub	moisture= 40% wt.	0.3	0.85	Jorgenson and Kreig, 1988
Tussock bog	moisture= 55% wt.	0.3	0.9	Jorgenson and Kreig, 1988
Aspen, spruce and hardwoods, feather moss	lacustrine silt	0.41	0.61	Taylor, 1995
Upland hardwood forest	organic silt	0.37	0.59	Taylor, 1995
Spruce forest, feather moss	lacustrine-eolian sand	0.43	0.66	Taylor, 1995
Upland hardwood and spruce forest	lacustrine silt	0.2	0.3	Taylor, 1995
Small bog	-	0.12	0.66	Taylor, 1995
Open succession vegetation of burn	silt	0.36	0.48	Taylor, 1995
Tall open spruce, hardwood forest	silt	0.48	0.74	Taylor, 1995

Table 2.3 continued

Site description	Soil Conditions	Nf	Nt	Reference
Tall open spruce, hardwood forest	silt	0.36	0.44	Taylor, 1995
Fen	-	0.29	0.26	Taylor, 1995
Tall spruce, birch and hardwoods	stony	0.31	0.59	Taylor, 1995
Open upland spruce and hardwood	organic and silt	0.3	0.64	Taylor, 1995
Open upland spruce and hardwood	silt veneer over stony colluvium	0.23	0.42	Taylor, 1995
Open black spruce and hardwoods with moss	till	0.13	0.34	Taylor, 1995
White spruce forest, thin moss layer	silt	0.35	0.5	Burn, 1998
Post forest fire, willow aspen, spruce saplings	silt	0.22	0.55	Burn, 1998
Moist Nonacidic Coastal Plain	Pergelic cryoborolls	-	0.82	Klene, 2001
Wet Nonacidic Coastal Plain	Histic pergelic cryaquolls	-	1	Klene, 2001
Moist Nonacidic Northern Foothills	Pergelic cryaquolls	-	0.95	Klene, 2001
Moist Acidic Northern Foothills	Pergelic cryaquepts	-	0.81	Klene <i>et al.</i> , 2001
Barren Southern Foothills	Gravel	-	1.25	Klene <i>et al.</i> , 2001
Water Track Southern Foothills	Pergelic cryaquepts	-	0.63	Klene <i>et al.</i> , 2001
Wet Acidic Southern Foothills	Histic pergelic cryaquolls	-	0.92	Klene <i>et al.</i> , 2001
Moist Acidic Northern Foothills	Pergelic cryaquepts	-	0.76	Klene <i>et al.</i> , 2001
Shrubland Northern Foothills	Pergelic cryothents	-	0.97	Klene <i>et al.</i> , 2001
Water Track Northern Foothills	Histic pergelic cryaquolls	-	0.73	Klene <i>et al.</i> , 2001

During the winter, the freezing n-factor is thought to be primarily controlled by snow conditions. Riseborough and Smith (1998) examined the effects of snow depth on the freezing n-factor through a simulation model. They found that the freezing n-factor declined with snow depth, but was also a function of mean annual air temperature. For a given snow depth, the freezing n-factor is lower when the mean annual air temperature is higher, due to the freezing back of a thicker active layer. Thus, snow depth has a greater impact on the surface temperature at high mean annual air temperatures. Snow conditions have not been included in site descriptions, presumably due to their inter-annual variability (Table 2.3).

Subsurface thermal conditions also affect the freezing n-factor because they control the thermal gradient (Burn, 1998). Where all other site variables are the same, a site with a thin active layer would have a higher freezing n-factor than a site without permafrost. Unfortunately, *a priori* knowledge of subsurface conditions is not normally available, and is after all, commonly the purpose of the modeling exercise. While moisture, shading, snow, and subsurface conditions have been identified as controlling the n-factor, the relative influence of each characteristic has yet to be determined.

2.4.3.2 Spatial and temporal variability of n-factors

The site characteristics that control the n-factor vary spatially and temporally, thus the n-factor is also variable in these dimensions. However, n-factors are usually calculated from air and surface temperature measured at one location over one season. To determine how representative an n-factor is of the relation between air and surface

temperature of an area over many seasons, the spatial and temporal variability must be examined.

Karunaratne and Burn (2003) examined freezing n-factors and snow accumulation patterns at several sites in southern Yukon Territory over three winters. They found that the freezing n-factor varies inter-annually depending on snow depth and the timing of the snow cover. At a spruce forest site with permafrost, the freezing n-factor ranged from 0.48 to 0.57 over three years, while at a non-permafrost meadow site the freezing n-factor ranged from 0.40 to 0.55.

Klene *et al.* (2001) examined the spatial and temporal variability of thawing n-factors within 1 ha plots of similar site conditions in the low arctic tundra. Bare rock and gravel surfaces were found to have the highest thawing n-factors (1.25), while water track sites had the lowest thawing n-factors (0.63 and 0.73). These results indicated that surface moisture is a major determinant of thawing n-factors in tundra environments. Temporal variability of the thawing n-factor was assessed using eight to nine point measurements of ground surface temperature within 1 ha plot over three years. The spatial variability among the point measurements was greater than the temporal variability experienced by a single measurement over three years.

The variability of the relation between air and surface temperature throughout the season has yet to be examined. Since the n-factor is always calculated on a seasonal basis it is unknown how representative the n-factor is of the relation between air and surface temperature throughout the season.

2.5 Concluding remarks

This review suggests that there is a need to determine the behaviour of n-factors. To determine the affect of surface and subsurface characteristics on the n-factor, the relation between air and surface temperature and the physical characteristics where the temperature measurements were made must be monitored both spatially and temporally. The Mayo Scientific Research Reserve, central Yukon Territory, is an ideal site to examine the behaviour of n-factors because several sites with different characteristics exist in close proximity.

CHAPTER THREE

STUDY AREA AND METHODS

3.1 Introduction

This study investigates variations in the relation between air and surface temperature with changes in site characteristics. Air and surface temperatures were measured and site characteristics were monitored for one year at five sites with different surface and subsurface conditions. The sites were located within the Mayo Scientific Research Reserve (MSRR), central Yukon Territory. This chapter describes the location and physical characteristics of the study area. Field methods and those used to derive the n-factor are also described.

3.2 Mayo Scientific Research Reserve

3.2.1 Location and geomorphic setting

Mayo, 63°36'N, 135°53'W, has an elevation of 495 m a.s.l. (Figure 3.1). The Scientific Research Reserve is situated on Stewart River, approximately 3 km upstream from Mayo (Figure 3.2). Mayo Airport, where meteorological measurements are taken, is 3 km north of the village and has a slightly higher elevation (504 m a.s.l.)

At the MSRR, Stewart River valley is approximately 5 km wide, and is bounded by forested hills, some of which crest at over 1200 m a.s.l. The valley bottom is flat, and in places is covered by silty clay sediments deposited in a glacial lake which filled Stewart River valley following McConnell glaciation, 13 600 years B.P. (Bostock, 1966; Hughes, 1983). Currently, these sediments are exposed on the southern bank of Stewart

Figure 3.1 Map of Yukon Territory.

Figure 3.2 Aerial photograph of the Mayo Scientific Research Reserve upstream from Mayo. Red squares indicate study sites. Parts of aerial photograph A27482-84 © 1989 Her Majesty the Queen in Right of Canada.

River within the MSRR. On the other side of the river, the glaciolacustrine sediments have been eroded and replaced with sandy alluvial sediments that form a point bar.

3.2.2 Climate

The climate of the Yukon is sub-arctic continental. Typically there are large annual, day-to-day, and daily ranges in air temperature, low to moderate atmospheric relative humidity, and moderate or light and irregular precipitation. Unlike arctic climates, intrusions of relatively mild air are experienced during winter (Wahl *et al.*, 1987).

Mayo lies within the *Central Yukon Basin* climatic division, which typically has the most continental climate in Yukon (Wahl *et al.*, 1987). At Mayo, the mean annual air temperature is -3.6°C , and the annual total precipitation is 318 mm (Environment Canada, 1993); however both are highly variable interannually and throughout the year. The mean wind speed at Mayo is 5.8 km/hr (Wahl *et al.*, 1987).

During summer, Mayo can experience warm spells with air temperatures reaching 36°C (Wahl *et al.*, 1987). The warmest month is July (Figure 3.3). Approximately two thirds of the annual precipitation arrives during summer and is often associated with convective rainstorms. More precipitation is received in July than any other month (Figure 3.3). The average wind speed for June, July and August is 22 km/hr (Yukon Fire Management, 2000).

At Mayo winters are long, cold and dry. Air temperatures have reached as low as -62°C (Wahl *et al.*, 1987). The coldest month is January (Figure 3.3). Precipitation is low in winter, usually less than 25 mm per month, and generally declines as the season

Figure 3.3 Mean monthly air temperature and precipitation at Mayo Airport, 1926-2000 (Meteorological Service of Canada, 2000).

progresses (Figure 3.3). Wind speeds are exceptionally low during winter allowing atmospheric inversions to develop in the valley (Burn, 1993).

3.2.3 Permafrost

Mayo lies within the widespread discontinuous permafrost zone where 50 to 90% of the terrain is underlain by permafrost (Heginbottom *et al.*, 1995). Within the MSRR permafrost conditions are variable and are predominantly dependent on the age of the sediment and surface characteristics. The undisturbed glaciolacustrine silty clay sediments on the south side of the river, contain ice-rich permafrost, and have a mean annual ground temperature that is approximately -1.3°C (Burn, 1992). Permafrost is approximately 34 m thick in this area (Burn, 1992). Within these sediments, thermokarst lakes (Burn and Smith, 1990) and retrogressive thaw slumps (Burn, 2000) are at varying stages of development and stabilization. Permafrost is degrading beneath active and recently stabilized thermokarst lakes and thaw slumps (Burn and Smith, 1990; Burn, 2000).

Permafrost is aggrading into the sandy alluvial sediments that form the point bar on the north side of the river. Permafrost is present in older alluvial sediments, but is thinner and warmer in younger material. Permafrost is absent from recently deposited alluvial sediments. Permafrost distribution along the point bar is similar to other point bars within the northern boreal forest (Viereck, 1970; Smith, 1975).

3.2.4 Soils

Eutric Brunisol and Turbic Cryosol are the dominant soil types in the region (White *et al.*, 1994). Eutric Brunisolic soils occur in forests on basic parent material, and lack fully developed leached and enriched surface horizons (Soil Classification Working Group, 1998). Turbic Cryosolic soils have permafrost within 2 metres of the surface, and have mineral horizons that have been cryoturbated (Soil Classification Working Group, 1998). Soil development is controlled by vegetation succession and permafrost conditions, which are dependent on the age of the sediment (Viereck, 1970).

Within the MSRR, soil conditions are dependent on the parent material and the stage of soil development. Three types of parent material are present within the MSRR; ice-rich glaciolacustrine silty clay, glaciolacustrine sediments that have been disturbed by retrogressive thaw slumps and thermokarst activity, and sandy alluvial sediments deposited by the river. Soil development is advanced on the undisturbed glaciolacustrine deposits. In these soils, organic surface horizons have formed and hummocks are present indicating permafrost and cryoturbation. Soil development on disturbed glaciolacustrine sediments depends on the time since the disturbance. Thermokarst disturbances result in organic-rich parent material (Smith, C.A.S., personal communication 2001), while thaw slumps result in organic poor parent material (Burn and Friele, 1989). The soils that have formed in the alluvial sands of the point bar are increasingly well developed as the distance from the current point of deposition increases. Young soils, formed on recently deposited sands, have no organic horizon, and are well drained, while older soils have a distinct organic horizon, and are poorly drained due to permafrost that has aggraded into the sediments.

3.2.3 Vegetation

The MSRR lies within the northern boreal forest. Vegetation distribution within the MSRR is associated with soil conditions (especially soil moisture and temperature), disturbance history, and channel migration. In stable terrain, white spruce (*Picea glauca*) is the dominant tree species where the soil is warm and dry (Oswald and Senyk, 1977). The herb and shrub layer in the white spruce forests is dominated by kinnikinnick (*Arctostaphylos uva-ursi*), wild rose (*Rosa acicularis*), and arctic lupin (*Lupinus arcticus*) (Cody, 1996). In wet areas, especially where permafrost is present, black spruce (*Picea mariana*) is the dominant tree species, and is often found with white birch (*Betula papyrifera*), labrador tea (*Ledum groenlandicum*), bog cranberry (*Vaccinium-vitis idaea*), and red bearberry (*Arctostaphylos ruba*) (Cody, 1996). In both black and white spruce forests the ground is often covered with feather mosses such as *Hylocomnium splendens*, and lichens such as *Cladina* spp. and *Cladonia* spp. (Cody, 1996).

In areas disturbed by retrogressive thaw slumps, the vegetation succession progresses from field horsetail (*Equisetum arvense*) and wild rhubarb (*Polygonum alaskanum*) to willow (*Salix* spp.), and then to mixed stands of willow/white spruce and willow/paper birch (Bartleman *et al.*, 2001). The vegetation succession is expressed spatially along the length of the slump with field horsetail dominating the vegetation near the recently stabilized headwall and spruce and birch trees present along the river bank (Burn, 2000).

Along the point bar, vegetation progresses from recently deposited bare sand to thickets of sandbar willow (*Salix exigua*), to thick stands of balsam poplar (*Populus balsamifera*) further inland. Past the balsam poplar, stands of white spruce and black

spruce have developed. The vegetation succession along the point bar is similar to that described by Viereck, (1970), and Smith, (1975).

3.3 Study period

Air and surface temperatures were measured during winter 2000 to 2001 and during summer 2001. Since n-factors are interannually variable due to variations in air temperature and precipitation, it is important to understand how representative the study period is of mean conditions. The study period extended from October 2000 to September 2001, and the air temperature and precipitation measured at the airport for the study period were compared with long-term records (Figure 3.4, and Figure 3.5)

The mean air temperature for the study period (-0.9°C) was 2.7°C warmer than the mean annual air temperature. This difference can mostly be attributed to the warm spell in January 2001 (Figure 3.4). The total precipitation for the study period was 366 mm; 42 mm higher than the mean annual total precipitation. During the study period, October, December and July received unusually high amounts of precipitation, while November and August received unusually low amounts of precipitation (Figure 3.5).

3.4 Field methods

3.4.1 Site selection

The MSRR is an ideal study area to examine the behaviour of n-factors for several reasons. First, many different permafrost conditions are represented within a small, easily accessible area. Second, meteorological measurements have been collected for over 75 years at the Mayo Airport, which provided contextual support for the research

Figure 3.4 Mean monthly air temperature (1926-2000) (Meteorological Service of Canada, 2000), and monthly mean air temperature during the study period at Mayo Airport.

Figure 3.5 Mean monthly precipitation (1926-2000) (Meteorological Service of Canada, 2000), and monthly precipitation during the study period at Mayo Airport.

and additional climate data. Third, environmental research has been ongoing for over 20 years at the MSRR, which provided contextual background for the study (Burn and Friele, 1989; Burn and Smith, 1990; Burn 2000). Finally, logistic support is available at the Village of Mayo, approximately 3 km from MSRR.

The five measurement sites studied in the MSRR represented a range of permafrost conditions typical of the discontinuous permafrost zone (Figure 3.2). The *Stable Permafrost Lacustrine* site (SPL) has stable permafrost in glaciolacustrine silty clay sediments. SPL was located in an open spruce forest amongst thermokarst lakes on the south side of Stewart River (Figure 3.2). The *Stable Permafrost Alluvial* site (SPA) had stable permafrost in alluvial sandy sediments. SPA was located in a closed spruce forest on the point bar to the north of Stewart River (Figure 3.2). The *Dry Permafrost Organic* site (DPO) had permafrost in dry organic-rich soil. DPO was located in a closed spruce forest amongst the thermokarst lakes to the south of Stewart River (Figure 3.2). The *Degrading Permafrost Slump* site (DPS) had degrading permafrost in silty clay thaw slump sediments. DPS was located amongst willows in the stabilized retrogressive thaw slump to the south of Stewart River (Figure 3.2). Finally, the *Non-Permafrost Alluvial* site (NPA) had sandy alluvial sediments, which were permafrost free. NPA was located within a mature balsam poplar stand on the point bar to the north of Stewart River (Figure 3.2). The air, surface and ground temperature records for NPA are missing from May 25 to July 10, 2001 due to flooding.

3.4.2 Air and surface temperature

To examine the behaviour of n-factors, air and surface temperature were measured at the study sites from October 2000 to September 2001. Air and surface temperature were measured every two hours using thermistor temperature sensors (Onset Computing, HOBOTM, model TMC6-HA), and logged by four-channel data loggers (Onset Computing, HOBOTM, model H08-006-04). The sensors had a range of -40°C to 100°C, an accuracy of $\pm 0.5^\circ\text{C}$ at 20°C, and a resolution of $\pm 0.41^\circ\text{C}$ at 20°C. The air and surface temperature sensors were installed in July 2000.

Air temperature was measured 1.5 m above the ground surface in a radiation shield to protect the sensor from incident radiation and precipitation. The surface temperature sensor was mounted on a wooden dowel and inserted into the ground. The surface temperature was measured approximately 2 cm beneath the surface. The surface was defined as the ground – air interface. Moss on the surface was considered part of the ground; therefore at sites with moss cover thermistor sensors were inserted 2 cm into the moss. The temperature of the ground 2 cm beneath the surface can be considerably lower than the surface temperature where the thermal conductivity is low.

3.4.3 Ground temperature

Ground temperatures were measured to determine the subsurface thermal regime at the sites. The ground temperature sensors were mounted onto the same wooden dowel as the surface temperature at 20 cm and 100 cm from the surface. The ground temperatures were measured using the same type of sensors and loggers used to measure air and surface temperature. The ground temperatures were measured every two hours

from October 2000 to September 2001. The ground temperature sensors were also installed in July 2000.

3.4.4 Vegetation

Vegetation composition and structure at the sites was described to determine associations between the n-factor and vegetation. At each site, the vegetation composition was established by compiling lists of vascular and non-vascular plants within a 10 x 10 m plot. The lists were then divided into three groups: trees, tall shrubs, and ground cover. Vegetation was described in August 2001.

3.4.4.1 Trees

This group contained tree species. Vegetation structure of the trees was documented by determining vegetation coverage, breast-height diameter, and tree height. To examine vegetation coverage of the trees, coverage of each tree species within the plot was estimated using a coverage class (Table 3.1). These coverage classes are a modification of the Braun-Blanquet cover scale (Goldsmith *et al.*, 1986).

The diameter at breast height (DBH) and the height of twenty trees closest to the air and surface temperature sensors were measured. The DBH was measured 1.5 m from the ground surface using a DBH measuring tape. Tree height was determined by measuring the angle between a horizontal plane and the tree crown from 10 m with a clinometer.

Table 3.1 Classes used to estimate vegetation cover (after Goldsmith *et al.*, 1986).

Class	Range of Cover
1	Present
2	< 1%
3	1 -5%
4	6 -15%
5	16 -30%
6	31 -50%
7	51 -75%
8	76 -100%

3.4.4.2 Tall shrubs

This group contained shrub species more than 0.5 m tall. Vegetation structure of the tall shrubs was described through vegetation coverage. To examine vegetation coverage of the tall shrubs, this group was subdivided into three height classes and a coverage class was assigned to each height class. The three height classes were defined as: >3 m, 1 to 3 m, and 0.5 to 1 m. The coverage classes described in Table 3.1 were used.

3.4.4.3 Ground cover

This group contained shrub species less than 0.5 m tall, herbaceous plants, mosses and lichens, as well as non-vegetative ground cover such as leaf litter and layering branches. Vegetation structure of the ground cover was explored through vegetation coverage. To examine vegetation coverage of the ground cover, a coverage class was assigned to each species. The coverage classes described in Table 3.1 were used.

3.4.5 Canopy cover

Canopy cover was estimated at each site several times throughout summer 2001 to determine the degree of (and changes in) shading produced by the vegetation. Hemispherical canopy photography was used to estimate canopy cover (Rich, 1990). Photographs were taken using a 180° hemispherical lens placed approximately 50 cm above the ground surface, within 2 m of the air and surface temperature sensors, under isotropic atmosphere conditions. The lens was directly perpendicular to the surface, and pointed towards the sky. The photographs were taken at each site on May 27, June 1,

July 4, and July 30 2001. The hemispherical photos produced circular images that show the dimensions and locations of gaps in the vegetation canopy. The images were analysed using Gap Light Analyser (Frazer *et al.*, 1999) to estimate canopy cover at the sites.

3.4.6 Rain

Daily precipitation was measured at each site from May 28 to August 15 to indicate periods of increased surface moisture availability and to determine differences among the sites in intercepted precipitation. Precipitation was measured daily with a manual rain gauge at each site.

3.4.7 Snow depth

Snow depth was measured to determine the insulating affect of snow cover at the sites. Snow depth was not measured during the study period, but rather in February 2002 (the following winter). Twenty snow depth measurements were made within 3 m of the air and surface temperature sensors using a graduated snow stake to determine variability of snow depth at each site. Since the snow depth measurements were taken outside the study period only the relative differences in snow depth among the sites can be inferred.

3.4.8 Surface soil moisture

Surface soil moisture content was measured to determine the degree of moisture availability at the sites. To measure surface soil moisture soil samples of a known volume were taken at a depth of 0 to 10 cm on August 9, 2001. The samples were dried

overnight in a soil oven at 105°C, and were weighed before and after drying to determine the loss of water from the samples. Gravimetric and volumetric water contents were determined for the samples as well as bulk density. The bulk density of the soil sample obtained at SPA was estimated from the literature, as the volume of that sample was unknown. In addition, daily observations of the presence of standing water and surface wetness at the sites were made throughout the summer.

3.4.9 Soil properties

The physical and thermal properties of the soil at each site were examined to determine the ability of the soil to transfer heat and water. To examine the soil properties soil pits were dug to a depth of 0.5 m at each of the sites in mid-July 2001. At 10 cm intervals from the surface, measurements of soil temperature, thermal conductivity, heat capacity, and thermal diffusivity were made using an ISOMET[®] needle probe (model 2104). Field descriptions of soil texture and organic content were also noted at 10 cm intervals.

3.4.10 Thaw depth

The thaw depth was measured several times throughout the summer 2001 at each site to determine the proximity of the frost table to the surface. The depth to permafrost was measured at the study sites by probing on August 19, 2001.

3.5 Calculation of n-factors

Freezing and thawing n-factors were calculated for each site. Freezing n-factors were calculated for winter 2000 – 2001, and thawing n-factors were calculated for summer 2001. The n-factors were calculated using freezing and thawing degree-days for the air and surface. The freezing and thawing seasons at each site were defined using surface temperature. Thus, the freezing season began when the daily mean surface temperature dropped below 0°C for three consecutive days, and ended when the daily mean surface temperature rose above 0°C for three consecutive days, and *vice versa* for the thawing season. The temperature on days within the freezing season when the air or surface temperature rose above 0°C was not subtracted from the freezing degree-day totals, and likewise days within the thawing season when the air or surface temperature dropped below 0°C were not included in the calculation of thawing degree-days. Finally, air and surface degree-day totals were derived for the seasons and used in equations 2.5 (n_f) and 2.6 (n_t). A thawing n-factor could not be calculated for NPA due to an incomplete temperature record.

CHAPTER FOUR

PHYSICAL CHARACTERISTICS OF THE STUDY SITES

4.1 Introduction

Exploration of n-factor behaviour involves examining the relation between air and surface temperature with respect to changes in surface and subsurface characteristics. This chapter describes the physical characteristics of the study sites

4.2 Vegetation

The vegetation composition and structure is variable among the study sites. The trees at the sites are summarized in Table 4.1, and the tall shrubs in Table 4.2. Short shrubs, herbs, mosses, lichens and non-vegetative cover are summarized in Table 4.3. Photographs of the study sites are given in Figure 4.1

The forest at SPL is composed of mature white and black spruce (*Picea mariana*, *P. glauca*) and paper birch (*Betula papyrifera*) trees that are several metres tall (Table 4.1). Tall willow shrubs (*Salix* spp.) are present in more open areas (Table 4.2). The forest floor is covered with many short shrubs, especially labrador tea (*Ledum groenlandicum*) and bog cranberry (*Vaccinium vitis-idaea*) (Table 4.3). Although many species of herbs are present at SPL the coverage is low (Table 4.3). The ground is entirely covered with mosses, especially feather moss (*Hylocomnium splendens*), as well as several lichens (*Cladina* and *Cladonia* spp.) (Table 4.3).

The forest at SPA is primarily composed of paper birch and white spruce trees (Table 4.1). Blue-green willow (*Salix glauca*) shrubs between 0.5 m and 3 m tall are

Table 4.1 Coverage class, mean diameter at breast height (DBH), and mean height of tree species at the study sites.

	SPECIES		COVERAGE*	SAMPLE SIZE	DBH (cm)	TREE HEIGHT (m)
	SCIENTIFIC NAME	COMMON NAME				
SPL	<i>Betula papyrifera</i>	paper birch	4	3	5	6
	<i>Picea glauca</i>	white spruce	4	7	13	7
	<i>Picea mariana</i>	black spruce	6	10	4	3
SPA	<i>Betula papyrifera</i>	paper birch	5	8	5	7
	<i>Picea glauca</i>	white spruce	7	11	9	5
	<i>Picea mariana</i>	black spruce	1	1	4	5
DPO	<i>Betula papyrifera</i>	paper birch	2	4	15	15
	<i>Picea glauca</i>	white spruce	8	16	11	9
DPS	<i>Betula papyrifera</i>	paper birch	2	2	<2	2
	<i>Picea glauca</i>	white spruce	3	16	<2	1
	<i>Populus balsamifera</i>	balsam poplar	1	2	<2	2
NPA	<i>Alnus rugosa</i>	alder	1	1	3	4
	<i>Picea glauca</i>	white spruce	4	6	9	6
	<i>Populus balsamifera</i>	balsam poplar	8	13	21	19

* coverage classes described in Table 3.1

Table 4.2 Coverage class and species of tall shrubs at the study sites.

	COVERAGE*		SPECIES PRESENT	
	HEIGHT CLASS	COVER CLASS	SCIENTIFIC NAME	COMMON NAME
SPL	>3m	1	<i>Salix glauca</i>	blue-green willow
	1m -3m	2	<i>Salix exigua</i>	sandbar willow
	0.5m-1m	2	<i>Salix scouleriana</i>	scoulers willow
SPA	>3m	0		
	1m -3m	4	<i>Salix glauca</i>	blue-green willow
	0.5m-1m	3		
DPO	>3m	6	<i>Salix arbusculoides</i>	little-tree willow
	1m -3m	1		
	0.5m-1m	1		
DPS	>3m	4	<i>Salix glauca</i>	blue-green willow
	1m -3m	7	<i>Salix exigua</i>	sandbar willow
	0.5m-1m	5	<i>Salix monticola</i>	mountain willow
NPA	>3m	0	<i>Cornus stolonifera</i>	dogwood
	1m -3m	2	<i>Viburnum edule</i>	high-bush cranberry
	0.5m-1m	2	<i>Salix exigua</i>	sandbar willow
			<i>Salix glauca</i>	blue-green willow

* coverage classes described in Table 3.1

Table 4.3 Coverage class for ground cover species at the study sites. Coverage classes described in Table 3.1.

SCIENTIFIC NAME SHORT SHRUBS	COMMON NAME	COVERAGE					
		SPL	SPA	DPO	DPS	NPA	
<i>Vaccinium vitis-idaea</i>	bog cranberry	6	7				
<i>Rosa acicularis</i>	prickly rose	3	3	6	3	4	
<i>Shepherdia canadensis</i>	soapberry	3		6	6	3	
<i>Ledum groenlandicum</i>	labrador tea	5	8				
<i>Arctostaphylos ruba</i>	red bearberry	3		3			
<i>Empetrum nigrum</i>	crowberry	3					
<i>Ribes triste</i>	wild red current		2				
HERBS							
<i>Geocaulon levidum</i>	bastard toad-flax	3	3				
<i>Pyrola secunda</i>	one-sided pyrola		2				
<i>Hedysarum alpinum</i>	liquorice root	2			3	4	
<i>Linnaea borealis americana</i>	twinflower	2		6	2		
<i>Epilobium angustifolium</i>	fireweed	3		3	2	2	
<i>Equisetum scirpoides</i>	dwarf scouring rush	2					
<i>Parnassia palustris</i>	marsh grass of parnassus				3		
<i>Equisetum arvense</i>	common horsetail		7		8	3	
<i>Erigeron acis</i>	fleabane daisy				2		
<i>Platanthera hyperborea</i>	northern green orchid				1		
<i>Aster puniceus</i>	purple-stemmed aster					4	
<i>Lupinus arcticus</i>	arctic lupin						
<i>Astragalus alpinus</i>	alpine milk-vetch			2	2		
<i>Achillea millefolium</i>	common yarrow			2			
<i>Mertensia paniculata</i>	tall lungwort	2	.	3	2		

Table 4.3 continued

MOSSES	SPL	SPA	DPO	DPS	NPA
<i>Hypnum splendens</i>	8	8	5		3
<i>Mesia uliginosa</i>	3				
<i>Ceratodon purpureus</i>				8	
<i>Ptilium crista-castrensis</i>					
<i>Dicranum polysetum</i>	2		2		
LICHENS	SPL	SPA	DPO	DPS	NPA
<i>Cladonia mitis</i>	6	5			
<i>Peltigera aphthosa</i>	2	3	3		
<i>Peltigera neopolydactyla</i>				3	
<i>Cladonia rangiferina</i>	3				
<i>Cladonia scabruiscula</i>	4	4			
<i>Cladonia uncialis</i>		2			
<i>Lepidozia reptans</i>	2	3			
<i>Cladonia stellaris</i>		3			
NON-VEGETATIVE COVER	SPL	SPA	DPO	DPS	NPA
Exposed mineral soil				2	3
Exposed organic soil			1		
Needle litter		2	6		2
Leaf litter		6	3	3	8
Layering branches		3	4	3	6

Figure 4.1a-c Photographs of study sites SPL, SPA, and DPO.

Figure 4.1d-e Photographs of study sites DPS and NPA..

common (Table 4.2). Labrador tea and bog cranberry are more prevalent at SPA than at SPL (Table 4.3). Common horsetail (*Equisetum arvense*) has high coverage at SPA (Table 4.3), while feather moss and several species of lichens cover the ground (Table 4.3).

The forest at DPO is composed of both mature white spruce trees and large little-tree willow (*Salix arbusculoides*) shrubs (Tables 4.1, 4.2). Many of the willows at this site are dead. The forest floor is covered with patches of prickly rose (*Rosa acicularis*), soapberry (*Shepherdia canadensis*), and twinflower (*Linnaea borealis americana*) (Table 4.3). The moss cover at DPO is discontinuous and includes feather moss and wavy dicranum (*Dicranum polysetum*) (Table 4.3). Large areas of the ground are covered with needle litter (Table 4.3).

The vegetation at DPS is recovering from thaw-slump disturbance, therefore the trees are still saplings. White spruce saplings are the most prevalent but balsam poplar (*Populus balsamifera*) and paper birch are also common (Table 4.1). Tall willow shrubs are abundant (Table 4.2), but there are few short shrubs at this site, however patches of soapberry are common (Table 4.3). Common horsetail is ubiquitous and dominates the herb layer at this site (Table 4.3). Much of the ground at DPS is covered with purple horn-tooth moss (*Ceratodon purpureus*) and leaf litter and branches from the willows. There are patches of dark mineral soil exposed at this site (Table 4.3).

The forest at NPA is mature balsam poplar, with some white spruce trees (Table 4.1). Several species of tall shrubs are present at NPA but do not have high coverage (Table 4.2). The short shrubs and herbs at this site include prickly rose, soapberry,

liquorice root (*Hedysarum alpinum*), and purple-stemmed aster (*Aster puniceus*). The short shrubs and herbs at NPA have low coverage (Table 4.3). The ground is predominantly covered with leaf litter and layering branches but there are also some large patches of exposed dark mineral soil (Table 4.3).

4.3 Canopy cover

Hemispherical photographs of the vegetation canopy taken on July 30, and the average canopy cover on July 4 and July 30 at the study sites are presented in Figure 4.2. Canopy cover was lowest at DPS (39%) and SPL (49%). Canopy cover was highest at DPO and NPA (76% and 77% respectively). Although the canopy cover at DPO and NPA were similar, the canopies were quite different. At NPA the tall deciduous trees resulted in numerous small gaps in the canopy, while at DPO the shorter coniferous trees resulted in fewer large gaps in the canopy (Figure 4.2). The canopy cover at SPA was 67%.

Figure 4.3 presents canopy cover at the study sites on selected days throughout the leaf-out period. The greatest change in canopy cover over time occurred at DPS and NPA. These changes are expected since the dominant vegetation at these sites is deciduous (Tables 4.1, 4.2). The canopy cover also changed at SPA over the season due to the prevalence of paper birch (Table 4.1). At SPL and DPO there was little change in canopy openness over the summer, due to the low amount of deciduous vegetation at these sites (Tables 4.1, 4.2).

Figure 4.2 Hemispherical photographs of the vegetation canopy at each study site, July 30. Mean canopy cover for July 4 and 30 indicated in the bottom right-hand corner.

Figure 4.3 Canopy cover at the study sites throughout the leaf-out period.

4.4 Rain

Total precipitation at the study sites from May 28 to August 15, 2001 is presented in Figure 4.4. Since the sites were located within a small area, differences in total precipitation measured were dominantly a function of precipitation interception. The total precipitation was similar at DPS and NPA, the two sites with deciduous trees, despite differences in canopy cover. The total precipitation at the sites with coniferous trees was negatively correlated with canopy cover. Coniferous trees are highly effective at intercepting precipitation due to the high surface area of the needles. At SPL where the canopy cover was low, total precipitation was high, and at DPO where the canopy cover was high, total precipitation was low.

4.5 Snow cover

Snow depths at the study sites in February 2002 are presented in Figure 4.5. The differences in snow cover among the sites can be attributed to precipitation interception rather than snow redistribution, because the sites are located within a small area and the wind speed during winter is low. Similar to total precipitation, snow depth at the coniferous study sites is negatively correlated with canopy cover. Of the coniferous sites DPO has the thinnest snow cover and highest canopy cover, while SPL has the thickest snow cover and the lowest canopy cover. At DPS and NPA, the primarily deciduous sites, snow cover was relatively high due to the high canopy openness during winter when the deciduous trees are bare. The snow depth was slightly greater at DPS where the coniferous saplings are too small to intercept snowfall. Except for DPO, the snow depths at the study sites were very similar.

Figure 4.4 Total precipitation at the study sites between May 28 and August 15, 2001.

Figure 4.5 Snow depth at the study sites, February 2002.

4.6 Surface soil moisture

The gravimetric surface (top 10 cm) soil moisture content at the study sites is reported in Table 4.4. These results indicate that SPA is the wettest site, while DPS is the driest. However, due to large variation in the bulk density of the surface soil among the sites (Table 4.4), due to differences in organic content, measurements of gravimetric soil moisture are not representative of soil moisture status. Measurements of volumetric soil moisture provide a better indication of surface wetness (Table 4.4). The surface volumetric soil moisture content was highest at DPS (0.64 g cm^{-3}) and NPA (0.60 g cm^{-3}). At DPO and SPA the surface volumetric soil moisture content was considerably lower (0.47 g cm^{-3} and 0.29 g cm^{-3} , respectively). The surface volumetric soil moisture content was lowest at SPL (0.21 g cm^{-3}).

Standing water was often observed at SPA and DPS following rainstorms. However at DPS the water remained for several days, while the moss surface at SPA quickly dried out. Although standing water was not observed at NPA (except when the site was flooded) the ground and leaf litter were usually wet to the touch for several days following rain events. Standing water was never observed at SPL or DPO. The ground at SPL was wet only immediately following rainstorms. At DPO the ground was often damp. These results indicate that DPS and NPA have relatively wet surfaces, DPO and SPA are drier, and the surface at SPL is dry.

4.7 Soil properties

The soils at the study sites have different physical characteristics, which result in different thermal properties. These differences in thermal properties among the study

Table 4.4 Surface soil moisture at the study sites on August 9, 2001.

	Gravimetric Soil Moisture (%)	Volumetric Soil Moisture (g cm ⁻³)	Bulk Density (g cm ⁻³)
SPL	45	0.21	0.46
SPA	244	0.29	0.12*
DPO	51	0.47	0.92
DPS	38	0.64	1.68
NPA	54	0.60	1.13

* estimated from Grigal *et al.* (1989)

sites both cause and result from different permafrost conditions. The soil physical and thermal properties of the study sites are summarized in Table 4.5.

At SPL the parent material of the soil is glaciolacustrine silty clay. The site is located on top of an earth hummock approximately 2 m across. A thin layer of mosses and undecomposed organic matter overlie a layer of semi-decomposed organics. Beneath the organic layer is silty clay of platy structure. This structure is indicative of the presence of segregated ice lenses in winter. The thermal properties at SPL differ greatly between the organic and mineral horizons (Table 4.5). Heat capacity (C) is highly dependent on moisture content and can therefore be used to indicate soil moisture. The heat capacity is low within the organic horizons, but is substantially greater in the silty-clay and increases with depth. The organic horizon has a low thermal conductivity (K) and thermal diffusivity (α), which results in a steep temperature gradient in midsummer. In contrast, the mineral horizon has a higher K and α and thus a moderate temperature gradient.

At SPA the parent material of the soil is alluvial sand. This site is located on uneven terrain with partially developed earth hummocks. A thick layer (~20 cm of mosses and undecomposed organic material overlies a thin horizon of semi-decomposed organic matter and silty sand, which in turn overlies a silty sand horizon (Table 4.5). The heat capacity is higher at the surface than at SPL but is similar at depth. The undecomposed organic matter has a low thermal diffusivity but greatly increases with depth as the amount of silty sand increases (Table 4.5). In midsummer, SPA has a gentler temperature gradient than SPL, and the soil temperature at a depth of 19 cm is cooler.

Table 4.5 Soil physical and thermal properties at study sites.

Site	Organic Thickness (cm)	Depth (cm)	Soil Description	T ^a (°C)	K ^b (Wm ⁻¹ K ⁻¹)	C ^c (x10 ⁻⁶) (Jm ⁻³ K ⁻¹)	α ^d (x10 ⁻⁶) (m ² s ⁻¹)
SPL	12	3	undecomposed organics	15.40	0.075	0.402	0.187
		9	semidecomposed organics	6.04	0.169	1.271	0.133
		19	silty clay	3.08	1.188	2.348	0.506
		29	silty clay	2.13	1.555	2.549	0.610
		39	silty clay	1.10	1.635	2.557	0.639
SPA	20	7	undecomposed organics	5.03	0.152	1.451	0.105
		19	undecomposed organics	1.68	0.439	2.009	0.219
		27	silty sand and semidecomposed organics	1.05	1.425	2.571	0.554
		32	silty sand	0.64	1.431	2.568	0.557
DPO	1	4	undecomposed organics	9.90	0.08	0.411	0.195
		14	semidecomposed organics, silt	7.40	0.497	2.049	0.243
		24	semidecomposed organics, silt	6.00	0.458	1.926	0.238
		34	semidecomposed organics, silt	4.60	0.432	1.842	0.235
		44	semidecomposed organics, silt	3.50	0.439	1.995	0.220
DPS	<1	1	silty clay, undecomposed organics	12.61	1.295	2.453	0.528
		8	silty clay	10.10	2.386	2.594	0.920
		18	silty clay	8.95	2.165	2.588	0.837
		28	silty clay	8.26	2.194	2.590	0.847
		38	silty clay	7.60	2.183	2.589	0.843
NPA	<1	5	silty sand and semidecomposed organics	11.96	0.456	2.516	0.181
		16	silty sand	10.47	0.635	2.524	0.252
		31	sand	9.88	1.429	2.539	0.563
		47	sand	9.29	2.144	2.955	0.726

^a T = ground temperature taken on June 26-27, ^b K = thermal conductivity, ^c C = heat capacity, ^d α = thermal diffusivity

At DPO the parent material of the soil has a high organic content and is possibly of thermokarst lake origin. This site is located on a flat surface on the top of a small hill. A thin moss layer (1 cm) overlies a layer of undecomposed organic matter. Beneath this layer the soil is composed of semi-decomposed organic matter and silt and is unvarying with depth (Table 4.5). The heat capacity at the surface is similar to the surface at SPL but is lower than any other site at depth. Low heat capacity is indicative of low soil moisture. The thermal diffusivity of the soil at DPO is lower than at any other site (Table 4.5). The temperature gradient at DPO is lower than in the organic horizons at SPL and SPA but higher than in the mineral horizons at these sites.

At DPS the soil has formed on silty-clay sediments but is not well developed. This site is located within a retrogressive thaw slump. The surface is covered with a thin layer (<1 cm) of discontinuous moss and undecomposed leaf litter. Beneath this layer the soil is silty clay, which is highly compacted, and unvarying with depth (Table 4.5). The heat capacity is high at this site and is generally unvarying with depth. The thermal diffusivity of the soil at DPS is higher than at any other site resulting in a low temperature gradient (Table 4.5).

At NPA the parent material of the soil is alluvial sand. Beneath the leaf litter that covers the ground is a layer of silty sand. The silty sand overlies sand which is unvarying with depth (Table 4.5). The thermal diffusivity at NPA is low near the surface, ($\sim 0.2 \times 10^6$) but increases with depth. This increase is possibly due to increased soil moisture content as indicated by the simultaneous increase in heat capacity with depth (Table 4.5). The thermal gradient at NPA is low and similar to the gradient experienced at DPS.

Although the soil thermal properties at the study sites were not examined during winter the thermal conductivity would be higher than in summer as the soil moisture freezes. However the relative increase in thermal diffusivity would depend on the thickness of the organic layer and seasonal changes in moisture conditions, and therefore vary significantly among the sites. For instance, thermal diffusivity at DPO would remain low during winter due to the low soil moisture content, and would increase at DPS due to the high soil moisture content. The greatest increase in thermal diffusivity during winter would occur at the sites with thick moss and peat layers.

4.8 Thaw depth

The rate of ground thawing varies among the study sites (Figure 4.6). The rate of thaw was highest at DPS due to the high thermal diffusivity of the ground at this site. In early summer the rates of thaw were similar at DPO and NPA, however by early June the thaw rate at NPA increased greatly. This increase was caused by a higher thermal diffusivity of the ground at depth at NPA. Throughout the summer, the thaw rates at DPO were slightly higher than at SPL because of the low soil moisture at DPO. The lowest thaw rate was experienced at SPA because of the low diffusivity of the ground at this site.

The seasonal thaw depth differs among the sites. At NPA, permafrost is absent. At DPS, permafrost is degrading, and the permafrost table is more than 3 m beneath the surface and was not reached by seasonal freezing (Burn, 2000). In late August 2001 thaw depth at SPL and SPA were 102 cm and 56 cm respectively, and was greater than 120 cm at DPO.

Figure 4.6 Thaw depths at the study sites during summer 2001.

4.9 Ground thermal regime

The ground thermal regime differs among the study sites, primarily due to differences in thermal diffusivity (Figure 4.7). In early October 2001, the ground temperature at a depth of 20 cm (T_{20}) decreased and approached 0 °C at all sites except SPA where it had already dropped below 0 °C (Figure 4.7). Between October and December, the sites experienced a zero curtain, with its length and timing varying among the sites. At SPA, the zero curtain period at T_{20} was longer than at any other site (75 days) due to the high soil moisture in the surface organic layer (Table 4.4, 4.5). At DPO, the zero curtain period was short (18 days) due to the low soil moisture at this site. Possibly the thin snow cover at DPO also contributed to the short zero curtain period. At SPL, the zero curtain period at T_{20} was longer (38 days) because of higher soil moisture at this site. At DPS and NPA, the zero curtain periods for T_{20} were long (54 and 44 days, respectively) and occurred later than at the other sites because of the higher surface heat capacity at these sites (Table 4.4, 4.5).

Following the zero curtain, T_{20} decreased at the study sites, but the rate of cooling varied considerably. The rate of cooling was highest at DPO possibly due to the high thermal gradient caused by a relatively thin snow cover. In addition, the cooling rate at DPO was not inhibited by latent heat released as the soil water froze because of the low soil moisture. At SPL and SPA the rate of freezing was lower than at DPO. These sites have higher soil moisture, therefore latent heat released at depth inhibits cooling at 20 cm. In mid-January, after the active layer was frozen at SPL and SPA, the cooling rate increased. At DPS and NPA the cooling rates were low throughout winter because the

Figure 4.7 Ground temperature at the study sites measured at (a) 20 cm and (b) 100 cm from the ground surface.

frost front did not reach a permafrost table, and therefore latent heat that was continually being released kept the soil relatively warm.

During winter T_{20} reached its minimum in late March at most sites. T_{20} was lowest at DPO during the first half of winter due to the short zero curtain period and high cooling rates experienced at this site. However T_{20} was lower at SPL during the second half of winter because of the higher thermal conductivity at this site once the active layer had frozen. Minimum T_{20} was warmer at SPA than at SPL. At DPS and NPA the minimum T_{20} were similar and warmer than at the other sites.

During April, T_{20} began to increase and approached 0°C , where it experienced the spring zero curtain period (Figure 4.7a). At all the sites except DPO the spring zero curtain periods were shorter than in fall and began at the same time. These seasonal differences are probably attributed to the melting of the snow cover. At DPO the closed canopy may protect the snow cover from melting and therefore prolong the snow cover at this site. This snow cover would have inhibited ground warming and resulted in a long (27 days) spring zero curtain period. Similar to the fall zero curtain period, the length of the spring zero curtain period was controlled by soil moisture. SPA had the longest zero curtain period (55 days), then NPA (39 days), and finally SPL (22 days). The spring zero curtain period at DPS was substantially shorter (14 days) than that experienced in the fall. This rapid warming was probably due to the high α at the near surface at this site (Table 4.5)

Following the zero curtain period T_{20} increased above 0°C (Figure 4.7a). The rate of increase varied among the sites and was primarily affected by subsurface conditions. Where permafrost is present warming rates were low as heat was used not only to warm

the ground but also to melt ice-rich soil at the base of the active layer. Thus, warming rates were highest at DPS where the permafrost is disconnected from the seasonally frozen soil, and at NPA where permafrost is absent. Warming rates for T_{20} were higher at DPO, than at SPL due to the low soil moisture at DPO. At SPA the warming rates were particularly low for T_{20} due to the low thermal conductivity of the overlying organic soil as well as the proximity of the permafrost table.

Maximum values of T_{20} are reached in July (Figure 4.7a). At DPS and NPA the maximum T_{20} are similar ($\sim 14^{\circ}\text{C}$) and are higher than the other sites. At DPO and SPL the maximum temperatures are cooler due to the presence of permafrost. DPO is warmer than SPL because of its low heat capacity. The maximum T_{20} at SPA during summer is lower than at any other site due the presence of permafrost and the thick overlying organic layer at this site.

The temperature at a depth of 100 cm (T_{100}) at the sites is warmer than T_{20} during winter and colder than T_{20} during summer (Figure 4.7b). Although the thermal regime at T_{100} is similar to T_{20} at the study sites, the effect of subsurface conditions is more pronounced at T_{100} . At SPA, T_{100} was recorded within permafrost, and did not exceed 0°C . T_{100} was measured slightly above the permafrost table at SPL, and rose above 0°C for a short period during summer. At SPL, T_{100} was warmer during summer and colder during winter than SPA due to the higher thermal conductivity at SPL. At DPO, T_{100} was warmer than at SPL or SPA in both winter and summer due to the deeper permafrost table at this site. At DPS and NPA, T_{100} were warmer than at the other sites due to the depth of the permafrost table (DPS) and the lack of permafrost (NPA).

4.10 Site Descriptions

Table 4.6 indicates the rank order of the study sites for the physical characteristics. Summary site descriptions follow:

SPL

SPL has stable permafrost in lacustrine sediments. This site is located on the top of a hummock in a spruce/birch forest. The canopy openness is high at this site. At SPL the ground surface is covered with feather mosses and lichens and the surface soil moisture is low. The thermal conductivity of the soil is low within the surface organic layer but increases substantially with depth. The permafrost table is 102 cm beneath the surface.

SPA

SPA has stable permafrost in alluvial sediments. This site is located within a spruce/birch forest. The canopy openness is low at this site. The ground at SPA is covered with a thick mat of mosses and undecomposed organic matter. The surface is wet at this site. The thermal conductivity at this site is low within the organic matter but increases with depth. The permafrost table is only 56 cm beneath the surface.

DPO

DPO has dry permafrost in sediments with high organic matter. This site is located on top of a small hill and is surrounded by spruce trees and large dead willow shrubs. The canopy openness is low at this site. Although the surface at this site is moderately wet, the underlying soil is dry, has a low thermal conductivity, and is generally unvarying with depth. The permafrost table at DPO is more than 120 cm from the surface.

Table 4.6 Ranking of study sites by physical characteristics. The same rank order is indicated by (=).

	July Canopy Cover	Snow Depth	Surface Wetness	Surface Organic Thickness	Sub- surface Heat Capacity (>20 cm)	Near- surface Thermal Diffusivity (<20 cm)	Sub- surface Thermal Diffusivity (>20 cm)	Depth to Permafrost	Winter Temperature at 100 cm depth	Summer Temperature at 100 cm depth
Highest	NPA ⁺	SPL ⁻	DPS	SPA	NPA	DPS	DPS	NPA	NPA	
	DPO ⁺	SPA ⁺	NPA	SPL	DPS	SPL	NPA	DPS	DPS	DPS
	SPA	DPS ⁺	DPO	DPO ⁺	SPL	DPO ⁺	SPL	DPO	DPO	DPO
	SPL	NPA ⁺	SPA	DPS ⁺	SPA	NPA ⁺	SPA	SPL	SPA	SPA ⁺
Lowest	DPS	DPO	SPL	NPA ⁺	DPO	SPA	DPO	SPA	SPL	SPL ⁺

DPS

DPS has degrading permafrost in sediments deposited by a retrogressive thaw slump. This site is located amongst willows and has high canopy openness. The surface has a thin discontinuous moss cover, and exposed mineral soil. The surface wetness at this site is high. The thermal conductivity is high at this site and is unvarying with depth. The permafrost table is more than 3 m beneath the surface.

NPA

Permafrost is not present in the alluvial sediments at NPA. This site is located within a balsam poplar forest and has low canopy openness. Leaf litter and layering branches cover the mineral soil at this site. The surface soil moisture is high at NPA. The thermal conductivity is high and gradually increases with depth at this site.

CHAPTER FIVE

AIR AND SURFACE TEMPERATURE AT THE STUDY SITES

5.1 Introduction

In Chapter Four it was determined that the study sites have variable physical characteristics. Since the physical characteristics of a site control its energy fluxes, the air and surface temperatures at the study sites are likely variable. The purpose of this chapter is to examine the air and surface temperatures measured at the study sites, and to present the freezing and thawing n-factors. The following research questions are addressed in this chapter:

1. What are the differences in air and surface temperature among the study sites?
2. What physical characteristics control the surface temperatures at the study sites?
3. What physical characteristics control the freezing and thawing n-factors at the study sites?

5.2 Air temperature

Air temperatures measured at the study sites between October 2000 and September 2001 were compared to the air temperature at Mayo Airport (YMA), a standard meteorological station (Figure 5.1). The range of mean air temperature for this period among the study sites and YMA was 1.7 °C, and the correlation coefficients (r) between YMA and the study sites were all greater than 0.99 ($p = 0.01$). Thus, the air temperature regimes at the study sites and YMA were similar.

Figure 5.1 Air temperature at Mayo Airport and at the study sites, October 2000 to September 2001.

Winter

In general the daily mean air temperature was below 0 °C from mid October to mid April at the study sites and YMA (Figure 5.1). The mean winter temperature, measured from October 1 to April 30, ranged from -10.1 °C at YMA to -12.0 °C at SPA (Table 5.1). Warm spells when the air temperature rose above 0 °C occurred in late November and several times in early March (Figure 5.1). Although the air temperature during January did not exceed 0 °C, monthly mean temperature was warmer than in the neighbouring months, and was uncharacteristically warm compared to previous years (Figure 3.4). A weeklong cold spell occurred during late March when air temperatures were more than 15 °C below the mean monthly temperature (-11 °C) (Figure 3.4, 5.1). The minimum winter air temperature, approximately -41 °C, occurred in mid December. Throughout winter the air temperature was highly variable from day-to-day (Figure 5.1; Table 5.1), probably due to the development and deterioration of temperature inversions and incursions of mild air (see section 3.2.2).

Summer

The daily mean air temperatures were above 0 °C from late April to late September (Figure 5.1). The mean summer temperature, measured from May 1 to September 30, ranged from 9.9 °C at SPA to 11.9 °C at YMA (Table 5.1). The maximum air temperature, approximately 21°C, was reached in mid July (Figure 5.1). The day-to-day variability of air temperature was lower during summer than during the previous winter (Table 5.1).

Table 5.1 Mean and standard deviation of air temperature measured at Mayo Airport and study sites for winter 2000-01 (October 2000 to April 2001) and summer 2001 (May to September 2001).

	YMA	SPL	SPA	DPO	DPS	NPA*
Winter 2000-01	Mean	-10.1	-11.6	-11.9	-11.5	-11.9
	Standard Deviation	9.8	10.1	10.0	10.2	10.0
Summer 2001	Mean	11.9	11.3	9.9	11.0	10.8
	Standard Deviation	4.5	5.1	5.3	5.0	4.9

* missing data from May 25 to July 10

5.2.1 Comparisons of air temperature

While this study computed n-factors from air temperatures measured at the study sites, in any regional analysis they would be applied using the air temperature measured at YMA. Therefore, the differences in air temperature between YMA and each of the study sites must be known to determine the error involved in applying the n-factor using air temperature measured at YMA. As well, the differences in air temperature among the study sites must be known to examine the behaviour of the n-factors at the study sites.

5.2.1.1 Mayo Airport

The differences between the daily mean air temperature at YMA and each of the study sites throughout the study period are presented in Figure 5.2. During winter, the difference between the air temperature at YMA and the study sites ranged from 4 °C below the air temperature at DPS, to nearly 12 °C above the air temperature at SPA. During summer, the difference between the air temperature at YMA and the study sites was less variable than during winter, and ranged from 3 °C below the air temperature at SPL, to 4 °C above the air temperature at SPA. For the majority of the study period the air temperature at YMA was higher than at the study sites (Figure 5.2).

The principal axis was computed to compare the differences in air temperature between YMA and the study sites. The principal axis is the appropriate regression line for two variables with equal error variance (Mark and Church, 1977). The principal axes between YMA (independent) and each of the study sites (dependent) were computed for winter and summer to describe the seasonal differences in air temperature between YMA and the study sites. The principal axes were similar for each of the study sites; therefore,

Figure 5.2 Difference in air temperature between YMA and each of the study sites.

only the data for air temperature at YMA and SPL are illustrated in Figure 5.3. The principal axes for YMA and the study sites in both seasons are summarized in Table 5.2. All the coefficients of determination (r^2) exceed 0.93 (Table 5.2). None of the slopes of the principal axes between YMA and the study sites in winter and summer are significantly different from 1.0 at the 95% confidence level (Church and Mark, 1980) (Table 5.2).

In both seasons, the Y-intercepts of the principal axes between YMA and the study sites are all less than and significantly different from 0.0 at the 95% confidence level (Table 5.2). In winter, the Y-intercept ranges from -1.24 at DPO, to -1.70 at NPA, (Table 5.2), thus the air temperature at YMA is between 1.2°C and 1.7°C warmer than the study sites. The difference in winter air temperature between YMA and the study sites could be caused by differences in elevation, due to the prevalence of atmospheric inversions (Wahl *et al.*, 1987). Unfortunately, although the elevation of YMA is known, the exact elevation of each of the study sites is not known, but is approximately 10 m lower than YMA. In summer, the Y-intercept ranges from -2.05 at NPA, to -2.27 at DPS, thus the air temperature at YMA is between 2.1°C and 2.3°C warmer than the study sites. The difference in summer air temperature between YMA and the study sites is likely caused by differences in vegetation. Unlike the study sites, the vegetation at YMA is restricted to short grasses. The negligible canopy cover at YMA may result in an elevated Q^* which would cause a higher air temperature (Roberts, 2002).

These results indicate that air temperature in natural environments differs from the air temperature at a standard meteorological station. Since, n-factors are currently computed using air temperature measured in natural environments, caution should be

Figure 5.3 Relations between air temperature at YMA and SPL for winter 2000-2001 and summer 2001.

Table 5.2 The principal axes of data for air temperature at Mayo Airport (independent) and each of the study sites (dependent).

	SPL	SPA	DPO	DPS	NPA*
r^2	0.96	0.96	0.96	0.96	0.95
Slope	1.026	1.034	1.019	1.044	1.016
Y- Intercept	-1.27	-1.61	-1.24	-1.36	-1.70
95% Confidence interval for Y-Intercept**	-1.03 to -1.51	-1.35 to -1.86	-1.01 to -1.47	-1.10 to -1.61	-1.44 to -1.96
Winter 2000-01					
r^2	0.94	0.94	0.94	0.93	0.96
Slope	1.137	1.071	1.105	1.096	1.091
Y- Intercept	-2.21	-2.14	-2.19	-2.27	-2.05
95% Confidence interval for Y-Intercept**	-2.67 to -1.77	-2.58 to -1.72	-2.64 to -1.76	-2.76 to -1.8	-2.41 to -1.71
Summer 2001					

* missing data from May 25 to July 10, ** Y-Intercept significantly different from 0.00 if confidence interval does not include 0.00

used when estimating surface temperatures using n-factors and air temperature measured at a standard meteorological station.

5.2.1.2 Study sites

The principal axes for data from each of the study sites were computed for winter and summer to determine the seasonal differences in air temperature between the study sites (Tables 5.3, 5.4). All relations between the study sites have r^2 values greater than 0.98. The slopes of all principal axes between the study sites for winter and summer data are not significantly different from 1.0 at the 95% confidence level.

Analysis of the 95% confidence intervals for the Y-intercepts shows that there are significant differences in air temperature between many of the study sites. However the differences in air temperature between the study sites are less than ± 0.5 °C in winter, and less than ± 0.3 °C in summer. Since the accuracy of the sensors is ± 0.5 °C (see section 3.4.2), the differences in air temperature between the study sites lie within the accuracy of the sensors. Therefore, the air temperatures at the study sites are similar.

5.3 Surface temperature

Ground surface temperatures measured at the study sites are presented in Figure 5.4. Unlike air temperature, surface temperatures were variable among the study sites. The mean surface temperature over the study period ranged from 0.8 °C at DPO to 4.1 °C at DPS. However, while the surface froze and thawed the surface temperatures at the study sites were similar.

Table 5.3 The principal axes of winter air temperature between each of the study sites.

r^2					
INDEPENDENT	DEPENDENT				
	SPL	SPA	DPO	DPS	NPA
SPL	-	1.00	1.00	1.00	1.00
SPA	-	-	1.00	1.00	0.99
DPO	-	-	-	1.00	0.99
DPS	-	-	-	-	0.99

SLOPE					
	SPL	SPA	DPO	DPS	NPA
SPL	-	1.007	0.993	1.017	0.991
SPA	-	-	0.986	1.010	0.984
DPO	-	-	-	0.976	0.998
DPS	-	-	-	-	0.974

Y-INTERCEPT*					
	SPL	SPA	DPO	DPS	NPA
SPL	-	-0.332	0.020	-0.074	-0.441
SPA	-	-	0.347	0.261	-0.115
DPO	-	-	-	0.092	-0.461
DPS	-	-	-	-	-0.370

* bold values indicate significance from 0.00

95% CONFIDENCE INTERVAL FOR Y-INTERCEPT**					
	SPL	SPA	DPO	DPS	NPA
SPL	-	-0.27 to -0.39	0.08 to -0.04	-0.02 to -0.12	-0.35 to -0.53
SPA	-	-	-0.42 to -0.28	-0.33 to -0.2	-0.01 to -0.22
DPO	-	-	-	-0.59 to -0.72	-0.35 to -0.57
DPS	-	-	-	-	-0.26 to -0.48

** Y-Intercept significantly different from 0.00 if confidence interval does not include 0.00

Table 5.4 The principal axes of summer air temperature between each of the study sites.

r^2

INDEPENDENT	DEPENDENT				
	SPL	SPA	DPO	DPS	NPA
SPL	-	0.989	0.990	0.988	0.991
SPA	-	-	0.993	0.990	0.994
DPO	-	-	-	0.989	0.994
DPS	-	-	-	-	0.990

SLOPE

	SPL	SPA	DPO	DPS	NPA
SPL	-	0.943	0.972	0.964	0.948
SPA	-	-	1.031	1.022	1.000
DPO	-	-	-	1.009	0.974
DPS	-	-	-	-	0.987

Y-INTERCEPT*

	SPL	SPA	DPO	DPS	NPA
SPL	-	-0.068	-0.046	-0.144	0.145
SPA	-	-	0.025	-0.074	0.203
DPO	-	-	-	0.099	0.154
DPS	-	-	-	-	0.254

* bold values indicate significance from 0.00

95% CONFIDENCE INTERVAL FOR Y-INTERCEPT**

	SPL	SPA	DPO	DPS	NPA
SPL	-	-0.22 to 0.08	-0.20 to 0.1	-0.30 to 0.01	0.00 to 0.29
SPA	-	-	-0.10 to 0.14	-0.22 to 0.07	0.08 to 0.32
DPO	-	-	-	-0.44 to -0.13	0.03 to 0.28
DPS	-	-	-	-	0.09 to 0.41

** Y-Intercept significantly different from 0.00 if confidence interval does not include 0.00

Figure 5.4 Surface temperature at the study sites, October 2000 to September 2001.

Table 5.5 Mean winter surface temperature and mean winter temperature at 100 cm at the study sites.

	SPL	SPA	DPO	DPS	NPA
Mean winter surface temperature (°C)	-3.5	-2.9	-5.2	-1.4	-1.6
Mean winter ground temperature at 100 cm (°C)	-2.2	-1.3	1.0	0.4	0.8

Fall and winter

In October, the surface temperatures were similar (slightly below 0 °C) at all study sites as the surfaces froze releasing latent heat (Figure 5.4). The surface temperatures at the study sites began to differ during November while the substrates froze. During winter the surface temperatures at the study sites were different (Figure 5.4). The coldest surface temperatures were experienced at DPO where surface cooling was not inhibited by latent heat released at depth because of the low soil moisture at this site. A thin snow cover at DPO may also have promoted surface cooling (Figure 4.5). The surface temperature at DPO increased more than at the other sites during the warm spells that occurred in early March (Figure 5.4), which also suggests DPO had a thinner snow cover, allowing the surface temperature at this site to respond quickly to changes in air temperature.

The warmest surface temperatures during winter were experienced at DPS and NPA (Figure 5.4). At these sites the freezing front did not reach a permafrost table, and therefore the surface at these sites was continually supplied with latent heat released at depth. The surface temperature at DPS was warmer during winter than at NPA perhaps due to a slightly thicker snow cover at DPS (Figure 4.5).

At SPL and SPA the surface temperatures were similar during early winter (Figure 5.4), due to their similar thermal diffusivities, heat capacities, and probable relative snow depths (Table 4.5 and Figure 4.5). However, in late January the surface temperature at SPL decreased more rapidly than at SPA. The increased cooling rate at SPL was caused by the freeze back of the active layer, which was completed in mid

January (Figure 4.7b). The active layer at SPA did not freeze back until the beginning of February (Figure 4.7b).

The rank order of the study sites by mean winter surface temperature is: DPO, SPL, SPA, NPA, and DPS. The rank order of the study sites by mean winter surface temperature is similar to the rank order by mean winter ground temperature at a depth of 100 cm, except for DPO (Table 5.5). The winter surface temperature at DPO was probably lower than the other sites, due to thinner snow cover at this site (Figure 4.5). The similarity in rank order between surface and subsurface winter temperature indicates that subsurface temperature is associated with surface temperature during winter. However, the considerable difference between DPO and the other study sites suggests that the effect of variable snow conditions exceeds the influence of subsurface effects.

Spring and summer

In early April, the surface temperatures at the study sites began to increase and become similar as the air temperature rose above 0 °C (Figure 5.4). During snowmelt, in late April, the surface temperatures at the study sites experienced a spring “zero curtain” that lasted approximately two weeks (Figure 5.4). During this period the surface temperatures at the study sites were all slightly below 0 °C and within 0.5 °C of each other.

Following this period, the surface temperatures at all the study sites increased (Figure 5.4). The rate of temperature increase varied among the sites, which resulted in different mean surface temperatures during the early summer (Table 5.6). The rank order of the study sites by mean early summer temperature (from May 1 to July 15) is: SPA,

Table 5.6 Mean early summer (May 1 to July 15) and summer (May 1 to September 30) surface temperature at the study sites.

	SPL	SPA	DPO	DPS	NPA*
Mean early summer surface temperature (°C)	9.31	5.86	7.82	12.16	N/A
Mean summer surface temperature (°C)	10.18	8.00	9.21	11.76	N/A

* missing data from May 25 to July 10

DPO, SPL, and DPS. Although NPA could not be ranked by early summer surface temperature due to missing data, the surface temperatures at NPA and DPO were similar until May 25 when the temperature record was interrupted (Figure 5.4).

The rank order of the study sites by early summer surface temperature is identical to the rank order by near-surface (top 20 cm) thermal diffusivity (Table 4.6). Thermal diffusivity controls the rate of thaw penetration. At DPS, where the near-surface thermal diffusivity is high, the frost table descended rapidly, whereas at SPA, where the near-surface thermal diffusivity is low, the lowering of the frost table was slower (Figure 4.6). The frost table affects the daily mean surface temperature when it lies within the zone of diurnal temperature fluctuation. Since the frost table is 0 °C, heat entering the ground during the day is used to thaw the ground rather than warm it, therefore the surface and subsurface temperatures are not in equilibrium on a daily basis. This imbalance between the surface and subsurface temperatures on a daily basis reduces the surface temperature. The influence of the frost table on the daily mean surface temperature diminishes as the frost table descends and is negligible when the frost table lies below the depth of daily temperature fluctuation.

The rank order of the study sites by early summer surface temperature is also similar to the rank order by canopy cover (Table 4.6). However, while NPA and DPO have similar early summer surface temperature (Figure 5.4) and similar canopy cover in July, during late May the canopy cover at the two sites is different. In late May, NPA, with a predominantly deciduous forest, had a canopy cover of 57%, while DPO, with a predominantly coniferous forest, had a canopy cover of 77% (Figure 4.3). Therefore canopy cover was not the primary determinant of early summer surface temperature at

the study sites. However, canopy cover is strongly correlated with Q^* (Roberts, 2002) and therefore probably contributed to the high early summer surface temperatures experienced at DPS and SPL where the canopy cover is low (Figure 4.2). Unfortunately the relative influence of canopy cover and near-surface thermal diffusivity on early summer surface temperature cannot be determined from this data set because they are correlated.

In June, the differences in surface temperature among the study sites began to decrease (Figure 5.4). By August, the daily mean surface temperatures at the study sites were similar, within 2 °C of each other, and continued so for the remainder of late summer and early fall. Since the canopy cover varied between the study sites during this period (Figure 4.3), the surface temperatures at the study sites are not controlled by canopy cover. The converging of the surface temperatures at the study sites coincided with the thawing of the ground below the maximum depth of diurnal temperature fluctuation at each of the sites. Based on the near-surface diffusivity (top 20 cm) the maximum depth of diurnal temperature fluctuation is approximately 42 cm at SPL, 33 cm at SPA, 39 cm at DPO, 74 cm at DPS, and 38 cm at NPA (Williams and Smith, 1989; equation 4.19). The thaw depth reached the maximum depth of diurnal temperature fluctuation in early June at SPL, DPO, NPA, and DPS, while at SPA this depth was not thawed until late June due to the low near-surface thermal diffusivity at this site. Therefore the similarity in surface temperatures among the sites in late summer and fall results from thawing of the ground below the maximum depth of diurnal temperature fluctuation, allowing the surface temperature to be dictated by air temperature, which was invariant among the study sites.

The rank order of the sites by mean summer surface temperature (from May 1 to September 30) is: DPS, SPL, DPO, and SPA (Table 5.6). Since the surface temperature at the study sites was the same during late summer, the early summer surface temperature dictated the rank order of the study sites for summer surface temperature. These results indicate that summer surface temperature is controlled by the rate of thaw penetration, which is dictated by near-surface thermal diffusivity.

5.4 Freezing and thawing n-factors

For each of the study sites, freezing n-factors were computed for winter 2000-2001, and thawing n-factors were computed for summer 2001. The n-factors were derived using air and surface freezing and thawing degree-days, and surface temperature was used to define the n-factor seasons. Table 5.7 presents the freezing and thawing periods, the air and surface freezing and thawing degree-days, and freezing and thawing n-factors for the study sites. Unfortunately, a thawing n-factor could not be calculated for NPA due to missing data.

The length of the freezing and thawing seasons, and the air freezing and thawing degree-days are similar at the study sites, while the surface freezing and thawing degree-days are different (Table 5.7). Therefore differences in the freezing and thawing n-factors among the study sites are due to differences in surface temperature.

The rank order of the study sites by the freezing n-factor is: DPS, NPA, SPA, SPL, and DPO. The rank order of the study sites by the freezing n-factor is reverse from the rank order by winter surface temperature. Therefore, low surface temperatures result in high freezing n-factors and *vice versa*. Since the winter surface temperatures

Table 5.7 Freezing and thawing n-factors for the study sites.

	SPL	SPA	DPO	DPS	NPA*
Freezing season 2000-01	Oct.1 to May 4	Sept. 29 to May 1	Oct.2 to May 6	Oct.1 to May 5	Oct.3 to May 11
FDDa	2524	2607	2489	2579	2559
FDDs	752	618	1106	309	351
Freezing n-factor	0.30	0.24	0.44	0.12	0.14
Thawing season 2001	May 5 to Oct. 6	May 2 to Oct. 6	May 7 to Oct. 6	May 6 to Oct. 10	N/A
TDDa	1738	1633	1676	1682	N/A
TDDs	1585	1231	1370	1808	N/A
Thawing n-factor	0.91	0.75	0.82	1.07	N/A

* missing data from May 25 to July 10

determine the differences in the freezing n-factor among the study sites the freezing n-factor must be controlled by snow depth and subsurface thermal conditions.

These results support the findings of Taylor (1995) who found that freezing n-factors were higher on exposed ridges where snow depths may be less. The influence of subsurface thermal conditions on the freezing n-factor identified in this thesis supports the findings of Riseborough and Smith (1998) who reported that the freezing n-factor was not only a function of snow depth but also mean annual air temperature. On a regional scale, subsurface thermal conditions such as active layer depth and ground temperature are controlled by mean annual air temperature.

The rank order of the study sites by the thawing n-factor is: SPA, DPO, SPL, and DPS. The rank order of the study sites by the thawing n-factor is identical to the rank order by summer surface temperature. Therefore, high surface temperatures result in high thawing n-factors and *vice versa*. Since the summer surface temperatures determine the differences in the thawing n-factor among the study sites, the thawing n-factor must be controlled by the near-surface thermal diffusivity, which controls the rate of thaw penetration.

Taylor (1995) reported that thawing n-factors were highest at open sites with good sun exposure and were lower at shaded sites. However, the other characteristics of these open sites indicate that they may also be associated with high near-surface thermal diffusivities; one site had exposed mineral soil, another had short grass ground cover, while the third was within an aspen forest with bog cranberry ground cover. On the other hand, the shaded sites were predominantly within spruce forests with thick moss covers, which are usually associated with low near-surface thermal diffusivity. Canopy openness

and near-surface thermal diffusivity may often be correlated, because low diffusivity mosses tend to develop in shaded areas where there is more moisture. However near-surface thermal diffusivity and not canopy cover controls the thawing n-factor as the surface temperatures at the study sites were similar following the descent of the frost table despite differences in canopy cover.

5.4.1 Potential variations of freezing and thawing n-factors

Since a standard method of computing n-factors has not been established within the literature, it is valuable to determine the effect of (1) using air temperature to define the freezing and thawing n-factor seasons, and (2) using air temperature measured at a standard meteorological site to compute the freezing and thawing n-factors. For each study site, Table 5.8 presents freezing and thawing n-factors computed with the freezing and thawing seasons defined by air temperature, and freezing and thawing n-factors computed using the air temperature measured at YMA.

The use of air temperature rather than surface temperature to define the n-factor seasons results in a decrease of 0.01 for the freezing n-factors, and a decrease of less than 0.03 for the thawing n-factors. The difference between the definitions is small because the magnitude of the air and surface temperatures at the beginning and end of the seasons are low and therefore do not greatly contribute to the degree-day totals. The use of air temperature measured at YMA rather than at the study sites to compute the n-factors resulted in an increase of 0.02 to 0.06 for the freezing n-factors, but a decrease of 0.05 to 0.11 for the thawing n-factors. The difference caused by using YMA air temperature is opposite for the freezing and thawing n-factors because the YMA air temperature is

Table 5.8 Freezing and thawing n-factors computed using air temperature to define the n-factor season and using air temperature measured at YMA.

	SPL	SPA	DPO	DPS	NPA*
Freezing n-factor	0.30	0.24	0.44	0.12	0.14
Freezing Season 2000-01					
Season defined by air temperature at study sites	0.29	0.23	0.43	0.11	0.13
YMA air temperature used to calculate n_f	0.34	0.28	0.50	0.14	0.16
Thawing n-factor	0.91	0.75	0.82	1.07	N/A
Thawing Season 2001					
Season defined by air temperature at study sites	0.88	0.74	0.82	1.07	N/A
YMA air temperature used to calculate n_t	0.85	0.67	0.78	0.99	N/A

* missing data from May 25 to July 10

greater at the study sites throughout the year. A warmer winter temperature decreases the total freezing degree-day for the air, while warmer summer temperature increases the total thawing degree-day for the air.

5.4.2 Principal axes of air and surface temperature

Burn (1998) proposed that the slope of the principal axis for air and surface temperature would be analogous to the n-factor if the Y-intercept were close to the origin. Air and surface temperature and the principal axis for their relations are plotted for each study site for the freezing season in Figure 5.5 and for the thawing season in Figure 5.6. Data for the principal axes analysis are presented in Table 5.9 for the freezing season and Table 5.10 for the thawing season.

The slopes of the principal axes for air and surface temperature at the study sites during winter are all lower than the freezing n-factors and the rank orders of the study sites by slope and by freezing n-factor are not the same (Table 5.9). Therefore the slopes of the principal axes do not approximate the freezing n-factors. The correlation coefficients (r) for the relations between air and surface temperature during winter (Table 5.9) are significantly different from 0.0 at the 95% confidence level at the study sites due to the large sample size ($n = 212$). However there is considerable scatter about the principal axis, which illustrates that the relation between air and surface temperature is variable on a daily basis.

The slope of the principal axis is closer to the freezing n-factor for DPO than for the other sites (Table 5.9). While the slopes for the other sites are an order of magnitude lower than the freezing n-factor, the slope for DPO is only 0.25 lower. The r for the

Figure 5.5 Principal axes for the relation between air and surface temperature for the freezing season at the study sites.

Figure 5.6 Principal axes for the relation between air and surface temperature for the thawing season at the study sites.

Table 5.9 Principal axes of air and surface temperature for the freezing season at the study sites.

Freezing Season					
	SPL	SPA	DPO	DPS	NPA
r	0.17	0.44	0.59	0.45	0.30
Slope	0.05	0.07	0.19	0.03	0.03
Y- Intercept	-3.18	-2.17	-3.25	-1.17	-1.39

Table 5.10 Principal axes of air and surface temperature for the thawing season at the study sites.

Thawing Season					
	SPL	SPA	DPO	DPS	NPA*
r	0.97	0.87	0.94	0.94	0.98
Slope	1.13	0.96	0.98	1.13	1.00
Y- Intercept	-2.57	-2.21	-1.50	-0.37	-0.36

*missing data from May 25 to July 10

relation between air and surface temperature during winter is also higher at DPO than at the other sites. The high slope and r at DPO indicate that the air and surface temperature are more closely related at DPO. These results support the hypothesis that the low surface temperatures at this site are caused by a thin snow cover.

The slopes of the principal axes for air and surface temperature at the study sites during summer are all higher than the thawing n -factors and the rank orders of the study sites by slope and by freezing n -factor are the same (Table 5.10). However the differences between the slopes of the principal axes and the thawing n -factors are variable among the sites. At SPA the slope of the principal axis is 0.21 higher than the thawing n -factor, while at DPS the slope of the principal axis is only 0.06 higher. These differences among the study sites suggest that the relation between air and surface temperature is variable among the study sites. The correlation coefficients (r) for the relations between air and surface temperature during summer (Table 5.10) are higher than during winter, and are significantly different from 0.0 at the 95% confidence level at the study sites, partially due to the large sample size ($n = 153$).

5.5 Summary

This chapter examined the differences in air temperature between the Mayo Airport and among the study sites, and the differences in surface temperature among the study sites. It was determined that the Mayo Airport was approximately 1.5 °C warmer than the study sites during winter, and was approximately 2 °C warmer during summer. The differences in air temperature among the study sites were less than ± 0.5 °C.

The mean winter surface temperatures at the study sites varied from -5.2°C at DPO to -1.4°C at DPS. All of the study sites have similar snow cover except for DPO where the snow cover is thinner. Since DPO has the coldest surface temperature, snow is the primary determinant of surface temperature. The rank order of the other study sites by winter surface temperature was identical to the rank order by mean annual ground temperature at 100 cm depth. Therefore where snow cover is uniform, subsurface conditions control the surface temperature during winter.

The mean summer surface temperature at the study sites varied from 8.0°C at SPA to 11.8°C at DPS. The surface temperatures at the study sites were variable during early summer and similar during the late summer and early fall. The rank order of the study sites by mean summer surface temperature at the study sites was identical to the rank order by near-surface thermal diffusivity. Thermal diffusivity controls the penetration of the frost table, which affects daily mean surface temperature when it intersects the diurnal temperature envelope. Therefore near-surface thermal diffusivity controls the surface temperature during summer.

The chapter concluded with a presentation of freezing and thawing n-factors. The freezing n-factor ranged from 0.12 at DPS to 0.44 at DPO while the thawing n-factors ranged from 0.75 at SPA to 1.07 at DPS. Since the differences in surface temperatures among the study sites exceeded those of air temperature, differences in freezing and thawing n-factors among the study sites were due to differences in surface temperature. The rank order of the study sites by freezing n-factor was identical to the rank order by winter surface temperature, and the rank order of the study sites by thawing n-factor was identical to the rank order by summer surface temperature. Therefore the freezing n-

factor is foremost controlled by snow cover conditions, and secondly subsurface thermal conditions. The thawing n-factor is controlled by near-surface thermal diffusivity through its effect on thaw penetration.

CHAPTER SIX

THE RELATION BETWEEN AIR AND SURFACE TEMPERATURE DURING THE FREEZING AND THAWING SEASONS

6.1 Introduction

The n-factor summarizes the relation between air and surface temperature on a seasonal basis. Chapter Five demonstrated that both air and surface temperature at the study sites were variable throughout the freezing and thawing seasons, suggesting that the relation between air and surface temperature may also vary through these seasons. The purpose of this chapter is to determine how representative the seasonal n-factor is of sub-seasonal conditions by examining the relation between air and surface temperature throughout the freezing and thawing seasons. The following research questions are addressed in this chapter:

1. What are the relations between air and surface temperature over the thawing and freezing seasons at the five study sites?
2. How representative is the n-factor of relations between air and surface temperature throughout the thawing and freezing seasons at the five study sites?

6.2 Freezing season

At the study sites, the freezing season extended from approximately the beginning of October 2000 to the end of April 2001 (Figure 6.1). During this period both the air and surface temperatures at the study sites were generally below 0 °C. At all of the study sites, the relations between air and surface temperature varied throughout the freezing

Figure 6.1 Air and surface temperatures at the study sites for the freezing season.

season. During October, air and surface temperatures at the study sites were close to 0 °C (Figure 6.1). However, the surface temperatures remained relatively constant, slightly below 0 °C, as they experienced the zero curtain, while the air temperatures fluctuated about the surface temperatures.

From November to March, surface temperatures were higher and less variable than air temperatures (Figure 6.1) due to the presence of snow. Snow inhibits the surface temperature response to changes in air temperature, therefore the relations between air and surface temperature constantly changed throughout the freezing season. When the air was relatively warm, as in January, the differences between air and surface temperature were only a few degrees Celsius. However, when the air was cold, as in mid-December, the differences between air and surface temperature were over 30 °C. At DPO, air temperature equaled and exceeded surface temperature several times during the freezing season because the surface was cold at this site for much of the season (Figure 5.3). A thin snow cover at DPO is thought to have caused the low surface temperatures, which resulted in a closer relation between air and surface temperature over the freezing season than at the other sites.

In April, air temperatures became less variable and were higher than surface temperatures, which were constant due to snowmelt (Figure 6.1). Thus, the differences between air and surface temperature were small and relatively constant in this period.

6.2.1 Seasonal evolution of the freezing n-factor

The changes in the relation between air and surface temperature over the freezing season are summarized by the freezing n-factor, using air and surface degree-days. Since

degree-days are cumulative over the season, seasonal evolution of the n-factor may be examined by plotting the daily cumulative ratio of surface to air freezing degree-days from the beginning of October to the end of April (Figure 6.2).

At the beginning of the freezing season the ratio may change rapidly from day-to-day because the degree-day totals are low at this time and therefore sensitive to daily additions. In October the ratio increased rapidly (Figure 6.2). The rapid increase in the ratio was caused by the relatively greater decrease in surface temperature than air temperature (Figure 6.1), which resulted in a relatively greater increase in surface freezing degree-days than air freezing degree-days.

After October, the daily cumulative ratios of surface to air freezing degree-days at SPA, DPS, and NPA declined as the relative decrease in air temperature exceeded those of surface temperature (Figure 6.2). At SPL the relative decrease in air and surface temperatures were similar, causing the ratio to remain fairly constant until mid-December. At DPO the relative decrease in surface temperature continued to exceed the relative decrease in air temperature causing the ratio to continually rise throughout the entire freezing season. By the end of December the daily cumulative ratios of surface to air freezing degree-days were less variable from day-to-day because the degree-day totals were large enough to be insensitive to daily additions.

At the end of December the ratios began to increase gradually (Figure 6.2) as the air temperatures rose (Figure 6.1). Towards the end of the season, the rate of increase in the ratios declined and eventually became constant as the daily additions were small, with the air and surface temperatures close to 0 °C. The final values of the daily cumulative ratio of surface to air freezing degree-days were the n-factors.

Figure 6.2 Daily cumulative ratios of surface to air freezing degree-days at the study sites.

The daily cumulative ratios of surface to air freezing degree-days illustrate how the relation between air and surface temperature varies over the freezing season. Towards the end of the season it is difficult to detect changes in the relation between air and surface temperature because the degree-day totals are too high to be sensitive to additions of low magnitude. Since the n-factor is a summation of temperatures, it is dominated by high-magnitude temperatures that occur during the middle of the freezing season. Therefore, the middle of the season, and the shoulder seasons are not equally represented by the n-factor. An alternative method must be used to examine the relation between air and surface temperature consistently throughout the freezing season.

6.2.2 Daily ratio of surface to air temperature

The relations between air and surface temperature during the freezing season varied on a daily basis at the study sites, due to the presence of snow and the highly variable air temperatures. To examine this daily variability throughout the freezing season, the daily ratio of surface to air temperature was plotted over this period for each of the study sites (Figure 6.3). Days when the air temperature exceeded 0 °C were not included because they are not included in the freezing n-factor.

During early/mid-freezing season, from October 1 to February 15, the daily ratios of surface to air temperature fluctuated about the seasonal freezing n-factors (Figure 6.3). When the air temperatures were high (low negative values), such as on November 22, the daily ratios were also high and exceed the seasonal freezing n-factors. Likewise, when the air temperatures were low (high negative values), the daily ratios were also low, and were less than the seasonal freezing n-factors. The mean daily ratios of surface to air

Figure 6.3 Daily ratios of surface to air temperature during the freezing season for the study sites.

temperature during early/mid-freezing season were similar to the seasonal freezing n-factors (Table 6.1). However, during this period, the standard deviations of the daily ratios of surface to air temperature were high, over half of the mean (Table 6.1). So, while the freezing n-factor represents the mean relation between air and surface temperature on a daily basis during early/mid-freezing season, the daily ratio of surface to air temperature is too variable to reliably estimate the seasonal freezing n-factor from a few daily measurements.

During the late freezing season, from February 16 to April 30, the daily ratios of surface to air temperature frequently exceeded the seasonal freezing n-factors and were highly variable from day-to-day, once the air temperatures rose above the surface temperatures (Figure 6.3). Similarly high ratios of surface to air temperature were not experienced during early winter because surface temperatures were only slightly less than 0 °C. Therefore when the air temperatures exceeded the surface temperatures they most likely exceeded 0 °C as well, and therefore were not included in this analysis or the n-factor calculation. The mean daily ratios during the late freezing season were higher than the seasonal freezing n-factors, so the freezing n-factor does not represent the relation between air and surface temperature at the end of the season.

6.2.3 Sub-seasonal relations between air and surface temperature

During the early/mid-freezing season, the daily ratio of surface to air temperature is too variable to detect sub-seasonal changes in the relation between air and surface temperature. To examine the sub-seasonal relations between air and surface temperature, “ten-day freezing n-factors” and “monthly freezing n-factors” were computed for each of

Table 6.1 The freezing n-factor and the mean and standard deviations of the daily ratios of surface to air temperature for early/mid freezing season (October 1 to February 15), and late freezing season (February 16 to April 30) for each of the study sites.

	SPL	SPA	DPO	DPS	NPA
Freezing n-factor	0.30	0.24	0.44	0.12	0.14
Early/mid freezing season					
Mean	0.24	0.23	0.46	0.13	0.13
Standard Deviation	0.16	0.15	0.37	0.11	0.16
Late freezing season					
Mean	2.23	1.61	1.63	0.65	0.78
Standard Deviation	2.72	4.35	1.58	0.88	0.88

the study sites (Figures 6.4 and 6.5). At the study sites, both the ten-day and monthly freezing n-factors were generally lower during the early freezing season, October to December, and higher during the late freezing season, March and April, than the seasonal freezing n-factors (Figures 6.4 and 6.5). During the early freezing season, the ten-day and monthly freezing n-factors were low because of the large relative difference between air and surface temperature during this period. At the end of the freezing season, the ten-day and monthly freezing n-factors were high because of the small relative difference between air and surface temperature during this period.

During the mid-freezing season, January and February, the ten-day and monthly freezing n-factors were similar to the seasonal freezing n-factors at all the study sites. The similarities between the seasonal freezing n-factor and the ten-day and monthly freezing n-factors during the mid-freezing season, especially during February, are expected as the air and surface temperatures during this time were low for an extended period and therefore contributed greatly to the seasonal air and surface degree-day totals.

The early, mid and late freezing season n-factors for the study sites are presented in Table 6.2. As expected the early freezing season n-factors were less than the seasonal freezing n-factors, the mid-freezing season n-factors were similar, and the late freezing season n-factors were greater than the seasonal freezing n-factors. Therefore the seasonal freezing n-factor is representative of the mid-freezing season relation between air and surface temperature but overestimates the ratio of surface to air temperature during the early freezing season and underestimates it during the late freezing season.

To determine the importance of the early and late freezing seasons to the seasonal freezing n-factors, the early to mid season freezing n-factors (October to February), and

Figure 6.4 "Ten-day freezing n-factors" for the study sites.

Figure 6.5 Monthly freezing n-factors at the study sites.

Table 6.2 The freezing n-factors and sub-seasonal freezing n-factors for the study sites.

	SPL	SPA	DPO	DPS	NPA
Freezing n-factor	0.30	0.24	0.44	0.12	0.14
Early season (October to December)	0.15	0.17	0.33	0.09	0.07
Mid season (January & February)	0.28	0.22	0.47	0.11	0.14
Late season (March & April)	0.88	0.54	0.75	0.27	0.34
Early to mid season (October to February)	0.21	0.19	0.39	0.10	0.10
Mid to late season (January to April)	0.43	0.30	0.54	0.15	0.19

the mid to late season freezing n-factors (February to April) were computed and are also presented in Table 6.2. Although the early to mid season n-factors were higher than the early season n-factors, and the mid to late season n-factors were lower than the late season n-factors, the early to mid season n-factors were lower and the mid to late season n-factors are higher than the seasonal freezing n-factors. Therefore, the early and late freezing seasons are important components of the freezing season; however, they have opposing effects on the seasonal freezing n-factor. Therefore, in determining the seasonal freezing n-factor, the early and late seasons must be either both excluded or both included.

6.3 Thawing season

At the study sites the thawing season extended from approximately the beginning of May to the end of September 2001 (Figure 6.6). During this period both the air and surface temperatures at the study sites were generally above 0 °C. The relations between air and surface temperature at the study sites were variable throughout the thawing season. At the beginning of the thawing season the surface temperatures were generally less than the air temperatures, while towards the late thawing season the surface temperatures became similar to the air temperatures (Figure 6.6). These changes in the relation between air and surface temperature over the thawing season are due to the descent of the frost table beneath the depth of diurnal temperature fluctuation. At the beginning of the thawing season the surface temperatures were less than the air temperatures because the frost tables intersected the diurnal temperature envelopes, keeping the surface temperatures low. At the end of the thawing season the air and

Figure 6.6 Air and surface temperatures at the study sites for the thawing season.

surface temperatures were similar because the surface temperatures were controlled primarily by the air temperatures. At this time the frost tables did not affect the surface temperatures because they were lower than the diurnal temperature envelopes.

The point at which the surface temperatures became similar to the air temperatures varied among the study sites due to differences in near-surface thermal diffusivity (Figure 6.6). At SPA, where the near-surface thermal diffusivity is low, the surface and air temperatures became similar at the end of July. At SPL and DPO, the surface temperatures equaled the air temperatures earlier than at SPA due to the higher near-surface thermal diffusivity at these sites. The surface and air temperatures became similar in mid-June at SPL, and in early July at DPO. At NPA, the surface temperature was lower than air temperature during May and similar to air temperature in July when the air and surface temperature records resumed. At DPS, the surface temperature became similar to the air temperature by mid-May, earlier than any other site, due to the high near-surface thermal diffusivity at this site (Figure 6.6).

During June and early July the surface temperature at DPS was higher than the air temperature (Figure 6.6). These high surface temperatures at DPS resulted from low Q_E . Q_E was low during this period because there was no standing water at DPS due to the low precipitation during June (Figure 3.5). In late July and early August, the surface temperature at DPS was lower, similar to the air temperature (Figure 6.6). The change in the relation between air and surface temperature was due to an increase in Q_E . During late July and early August, Q_E was higher because there was standing water at DPS from the precipitation received during July (Figure 3.5). In late August and early September, the surface temperature at DPS was again slightly higher than the air temperature (Figure

6.6) due to the low precipitation received during August (Figure 3.5). Although there was less precipitation received in August than June, the surface temperature in August was only slightly warmer than the air temperature. This difference between June and August in the relation between air and surface temperature was probably because the day length was approximately four hours shorter during August, causing $K_{\downarrow}$ to be lower during this period. During September, precipitation was higher than in June or August (Figure 3.5), which resulted in similar air and surface temperatures due to an elevated Q_E (Figure 6.6).

The association between surface wetness and the relation between air and surface temperature was observed following the retreat of the frost table to below the diurnal temperature envelope at DPS and to a lesser extent at NPA but not at the other study sites. DPS and NPA have the highest surface wetness (Table 4.4), and are the only sites with dark exposed mineral soil (Table 4.3). Since dark soils have low albedos (Table 2.1), the surface albedos at DPS and NPA are probably lower than at the other study sites. Therefore, when the surfaces at DPS and NPA were dry, Q_H was high because $K_{\uparrow}$ was low, which resulted in high surface temperatures. Conversely, when the surfaces at DPS and NPA were wet, Q_E was high because of the high moisture availability, which resulted in a relatively low surface temperature equal to the air temperature. The other study sites do not have low enough surface albedos to cause surface temperatures that exceed the air temperatures during dry periods, or wet enough surfaces to reduce the surface temperature during periods of high precipitation.

6.3.1 Seasonal evolution of the thawing n-factor

The thawing n-factor summarizes the various relations between air and surface temperature throughout the thawing season. To examine the seasonal evolution of the thawing n-factors the daily cumulative ratio of surface to air thawing degree-days for each of the study sites is shown in Figure 6.7.

At the beginning of the thawing season, the daily cumulative ratio increased at all the study sites because the relative increases in surface temperatures exceeded those of the air temperatures (Figure 6.7). Since the rate of surface warming varied among the study sites, the rate of increase in the daily cumulative ratio was also variable among the sites. At DPS, the cumulative ratio of surface to air degree-days increased rapidly because the high near-surface thermal diffusivity at this site caused the frost table to descend rapidly which resulted in the increase of surface temperature. Conversely, at SPA the cumulative ratio increased slowly because the low near-surface thermal diffusivity at this site inhibited the thawing of the frozen ground, which caused low surface temperatures. At SPL, the cumulative ratio increased faster than at SPA but slower than at DPS. At DPO, the rate of increase was initially faster than at SPL, but by the beginning of June, the rate decreased significantly. Although the cumulative ratio of surface to air thawing degree-days could only be plotted until mid-May at NPA, the rate of increase was similar to that of DPO, which is to be expected due to their similar near-surface thermal diffusivities (Table 4.5).

The increase of the daily cumulative ratios of surface to air thawing degree-days began to decline sometime after May, and the ratios eventually became fairly constant (Figure 6.7). The cumulative ratios became constant because the degree-day totals were

Figure 6.7 Daily cumulative ratios of surface to air thawing degree-days at the study sites.

large enough to be insensitive to daily additions. The final values of the daily cumulative ratios of surface to air thawing degree-days were the thawing n-factors.

The daily cumulative ratios of surface to air thawing degree-days illustrate how the relations between air and surface temperature evolve over the early thawing season. However, towards the end of the thawing season it was difficult to detect changes in the relation between air and surface temperature because the sensitivity of the cumulative ratio progressively decreased over time as the degree-day totals increased. Therefore it is difficult to determine the point at which surface and air temperatures become equal.

6.3.2 Daily ratio of surface to air temperature

To explore the relation between air and surface temperature throughout the thawing season, the daily ratios of surface to air temperature were plotted for each of the study sites (Figure 6.8). Days when the air or surface temperatures were less than 0 °C were not included. Based on the daily ratios of surface to air temperature, the thawing season was divided into three sub-seasons: the early thawing season when the daily ratios progressively increased, approaching 1.0; the mid thawing season when the ratio was fluctuating about 1.0; and the late thawing season when the ratio was variable.

The timing and duration of the early and mid thawing seasons were highly variable among the study sites, due to differences in near-surface thermal diffusivity (Figure 6.8). At SPL, the early season extended from early May to mid-June, and the mid season extended from mid-June to early September. At SPA, the early season extended from early May to the end of July, and the mid season extended from the end of July to early September. At DPO, the early season extended from early May to early July, and

Figure 6.8 Daily ratios of surface to air temperature during the thawing season for the study sites.

the mid season extended from early July to early September. At DPS, the early season extended from early May to mid-May, and the mid season extended from mid-May to early September. At NPA, the early season extended from early May to at least mid-May, and the mid season extended from at least mid July to early September. The late season extended from early September until the end of the thawing season at all the study sites (Figure 6.8).

At the beginning of the early thawing season the n-factors were greater than the daily ratios of surface to air temperature. Towards the end of the early season the daily ratios of surface to air temperature were similar to the n-factors. During the mid thawing season the daily ratios were generally slightly greater than the n-factors but less than 1.0. In the late thawing season the daily ratios were highly variable, and their relation to the n-factor were site specific; i.e., at SPL and SPA the daily ratios were less than the n-factor, while at DPO and DPS they were both greater than and less than the n-factor. Therefore, only at the end of the early thawing season were the thawing n-factors representative of the relations between air and surface temperatures at all the study sites.

The thawing n-factor is generally higher than the freezing n-factor because the difference between air and surface temperatures are less during the summer than in winter. Like the freezing n-factor, the thawing n-factor is controlled by the relation between air and surface temperature during mid season, rather than May or September, because the temperatures are high during these months and therefore greatly contribute to the seasonal degree-day totals. Therefore, the thawing n-factor is a function of the length of the mid thawing season. When the mid season is long, the air and surface temperatures are similar for the majority of the thawing season, and therefore, the

thawing n-factor is close to 1.0. When the mid season is short, the surface temperature is less than the air temperature for much of the thawing season, and therefore the thawing n-factor is less than 1.0. For example, at SPL where the mid season was long, the thawing n-factor was high, 0.91, and for most of the mid season the daily ratios of surface to air temperature were within 0.10 of the n-factor. Conversely, at SPA where the mid season was short, the thawing n-factor was low, 0.75, and during the mid season the daily ratios of surface to air temperature were higher than the thawing n-factor.

6.3.3 Sub-seasonal relations between air and surface temperature

The ten-day and monthly thawing n-factors at the study sites show the same trends throughout the thawing season as the daily ratios of surface to air temperature. At the study sites the ten-day thawing n-factors were generally lower than the seasonal thawing n-factors at the beginning of the early thawing season, and higher than the seasonal thawing n-factors at the end of the early thawing season and during the mid thawing season (Figure 6.9). Likewise, the monthly thawing n-factors at the study sites were generally lower than the seasonal thawing n-factors at the beginning of the early thawing season, and higher than the seasonal thawing n-factors at the end of the early thawing season and during the mid thawing season (Figure 6.10).

During the early thawing season the ten-day and monthly thawing n-factors were less than the seasonal thawing n-factors because the surface temperatures were lower than the air temperatures. During the mid thawing season the ten-day and monthly thawing n-factors were higher than the seasonal thawing n-factor because the air and surface

Figure 6.9 "Ten-day" thawing n-factors for the study sites.

Figure 6.10 Monthly thawing n-factors at the study sites.

temperatures were similar. The ten-day and monthly thawing n-factors were similar to the seasonal thawing n-factor at the end of the early thawing season.

To determine the importance of the early thawing season to the seasonal thawing n-factors at the study sites, n-factors were calculated for seasons that began progressively later from the beginning of the normal thawing season (Table 6.3). At SPL and DPS, the exclusion of May, June and even July resulted in values similar to the seasonal thawing n-factors. Therefore, at SPL and DPS, the early thawing seasons did not have large impacts on the seasonal thawing n-factors, because the early seasons were relatively short due to high near-surface thermal diffusivities. Conversely, at SPA and DPO, the exclusion of May, June and July resulted in progressively higher values than the seasonal thawing n-factors. Therefore, at SPA and DPO, the early thawing seasons had large impacts on the seasonal thawing n-factors, because the early seasons were relatively long due to low near-surface thermal diffusivities.

To determine the importance of the late thawing season to the seasonal thawing n-factors at the study sites, n-factors were calculated for seasons that excluded September. The n-factors produced from excluding September were: 0.91 at SPL, 0.73 at SPA, 0.82 at DPO, and 1.09 at DPS. The n-factors that excluded September were very similar to the seasonal thawing n-factors at all the study sites, therefore the late thawing season is unimportant to the seasonal thawing n-factor.

6.4 Summary

This chapter examined the relation between air and surface temperature throughout the freezing and thawing season to determine how representative the freezing

Table 6.3 The thawing n-factors and sub-seasonal thawing n-factors for the study sites with the sub season beginning on date specified and ending in early October.

	SPL	SPA	DPO	DPS	NPA *
Thawing n-factor	0.91	0.75	0.82	1.07	N/A
June 1	0.95	0.80	0.89	1.10	N/A
July 1	0.95	0.90	0.93	1.06	N/A
August 1	0.93	0.96	0.93	1.03	N/A
September 1	5.90	6.60	6.34	4.04	N/A

* missing data from May 25 to July 10

and thawing n-factors were of sub-seasonal conditions. For the freezing season, it was determined that the daily ratios of surface to air temperature from October to mid-February were variable but fluctuated about the seasonal n-factor value at all the study sites. After mid-February the daily ratios were higher than the seasonal n-factor and highly variable. Sub-seasonal freezing n-factors indicated that the seasonal freezing n-factor is representative of the mid freezing season relation between air and surface temperature but overestimates the ratio of surface to air temperature during the early freezing season and underestimates it during the late freezing season. It was also determined that due to the opposing nature of the relation between air and surface temperature during early and late freezing seasons, they must be either both included or both excluded when determining the seasonal freezing n-factor.

For the thawing season, it was determined that the daily ratios of surface to air temperature progressively increased towards 1.0 during the early thawing season, fluctuated between 1.0 and the seasonal thawing n-factor during the mid thawing season, and were highly variable during the late thawing season. The only period when the thawing n-factor well represents the relation between air and surface temperature at all the study sites is at the end of the early thawing season. The thawing n-factor is a function of the mid thawing season length, which was different among the study sites depending on the near-surface thermal diffusivity.

CHAPTER SEVEN

SUMMARY AND CONCLUSIONS

7.1 Summary of results

This thesis examines the behaviour of freezing and thawing n-factors at five study sites in the discontinuous permafrost of central Yukon Territory. Analysis of air and surface temperature measurements for one year determined that among the study sites surface temperature was variable and air temperature was not. The differences in surface temperature among the study sites resulted in different freezing and thawing n-factors. This thesis also determined that the relation between air and surface temperature varied throughout the freezing and thawing seasons. This chapter summarizes the behaviour of freezing and thawing n-factors, and presents conclusions of this thesis. Finally, some recommendations for future research are outlined.

7.1.1 The freezing n-factor

The freezing n-factor ranged from 0.12 to 0.44 at the study sites. Snow cover was identified as primarily controlling the freezing n-factor. A thin snow cover resulted in low surface temperatures that were related to air temperatures, while thicker snow covers caused higher surface temperatures, insulated from the air. Therefore, a thin snow cover resulted in a high freezing n-factor (0.44), while thick snow covers caused lower freezing n-factors (0.12 to 0.30). It was also determined that subsurface thermal conditions controlled the freezing n-factor when the snow covers were similar. Where permafrost was absent, latent heat released at depth throughout the freezing season kept the surface

warm. Where permafrost was present, latent heat was released only at the beginning of the freezing season, during freezing of the active layer, and subsequently surface temperatures decreased. Therefore, where snow covers were similar, unfrozen subsurfaces resulted in low freezing n-factors (0.12 to 0.14), while frozen subsurfaces were associated with relatively high freezing n-factors (0.24 to 0.30).

The relation between air and surface temperature throughout the freezing season was variable from day-to-day because of the snow cover, which isolated the surface from the air, which fluctuated constantly. Examination of sub-seasonal freezing n-factors determined that the freezing n-factor was representative of the mid freezing season relation between air and surface temperature but not the early and late freezing seasons.

7.1.2 The thawing n-factor

The thawing n-factor ranged from 0.75 to 1.07 at the study sites. Near-surface thermal diffusivity was identified as primarily controlling the thawing n-factor. Near-surface thermal diffusivity controls the lowering of the frost table, which inhibits surface warming when it intersects the diurnal temperature envelope. Therefore, high near-surface thermal diffusivity resulted in high thawing n-factors (0.91 to 1.07), while low near-surface thermal diffusivity resulted in low thawing n-factors (0.75 to 0.82). It was also suggested that surface albedo, and surface wetness affect the thawing n-factor. Where the surface was dry and had low albedo surface temperatures exceeded air temperatures, which resulted in thawing n-factors greater than 1.00.

The relation between air and surface temperature throughout the thawing season was variable among the study sites. Following the descent of the frost table below the

depth of diurnal fluctuation, air and surface temperatures were similar, whereas previously the surface temperatures were less than the air temperatures. Therefore, the point at which the surface temperature equaled the air temperature varied among the study sites depending on near-surface thermal diffusivity. The thawing n-factor was representative of the relation between air and surface temperature for the majority of the thawing season where the near-surface thermal diffusivity was high because the n-factor and the ratios of surface to air temperature were both close to 1.0. Where the near-surface thermal diffusivity was low the thawing n-factor was not representative of the relation between air and surface temperature for much of the season because the thawing n-factor was low and the ratios of surface to air temperature were progressively approaching 1.0.

7.2 Conclusions

Four conclusion regarding n-factor behaviour can be drawn from the examination of air and surface temperatures presented in this thesis:

- (1) Freezing n-factors are controlled primarily by snow cover but also by subsurface thermal conditions. Therefore snow and subsurface thermal conditions must be included in the detailed site descriptions that accompany freezing n-factors.
- (2) Thawing n-factors are controlled primarily by near-surface thermal diffusivity and therefore must be included in the detailed site descriptions that accompany thawing n-factors.

- (3) Freezing n-factors are representative of the relation between air and surface temperature only during the mid-freezing season, not on a day-to-day basis. The relations between air and surface temperature during the early and late freezing seasons are reverse and therefore must be either both excluded or both included.
- (4) Thawing n-factors are representative of the relation between air and surface temperature throughout much of the season only when the near-surface thermal diffusivity is high.

7.3 Recommendations for future research

Future research on the freezing n-factor should investigate the relative influence of snow and subsurface thermal conditions on surface temperatures under controlled conditions. To investigate the influence of snow cover on the freezing n-factor, snow manipulation experiments should be conducted within an area with homogeneous surface and subsurface conditions. To investigate the influence of subsurface thermal conditions on the freezing n-factor surface temperatures could be measured along an aggrading or degrading permafrost transect such as observed along an alluvial point bar (Smith, 1975) or a stabilized retrogressive thaw slump (Burn, 2000) where snow conditions are similar.

Future research on the thawing n-factor should investigate the influence of surface variables such as surface wetness, surface albedo, canopy cover etc., on surface temperatures. Since near-surface thermal diffusivity is the primary controller of surface temperatures during the thawing season to determine the influence of other surface

variables site manipulation experiments should be conducted under controlled near-surface thermal diffusivity conditions.

Finally, the relation between air and surface temperature throughout the freezing and thawing seasons should be examined over multiple years to determine long-term patterns and seasonal anomalies in the behaviour of the n-factor throughout the freezing and thawing seasons. Such an investigation would improve the confidence of the conclusions presented in this thesis.

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