THE DEVELOPMENT OF NEAR-SURFACE GROUND ICE AT ILLISARVIK, RICHARDS ISLAND, NORTHWEST TERRITORIES

By

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ABSTRACT

Samples of near-surface permafrost were obtained in summer 2010 from 26 sites in the Illisarvik drained lake basin, three sites on lake basin terraces, and nine sites in the surrounding tundra on Richards Island, NWT, to investigate the growth of near-surface ground ice and the relations between controlling factors and ice formation. Active-layer records from the basin, tundra, and lake terrace were also examined. Active-layer thicknesses were positively associated with summer air temperatures, and in some lake basin areas, with winter snow depths. Excess-ice accumulation was highly variable in the basin and predominantly positively associated with active-layer moisture status, and, to a lesser extent, soil texture. Near-surface excess-ice contents in the tundra were much greater than in the basin, and ice enrichment was present at greater depths, likely a result of permafrost aggradation since the early-Holocene climatic optimum.

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1. OVERVIEW AND OBJECTIVES

1.1 Introduction

This thesis examines the development of segregated ground ice at the top of permafrost. Segregated ice forms from the transfer of liquid water into freezing ground due to thermally induced pressure gradients (Mackay, 1972). Near-surface ground ice is widespread in permafrost regions with fine-grained soil, so that the top of permafrost is characteristically ice rich (Mackay, 1972). The field investigations were conducted at the Illisarvik experimental drained lake site on Richards Island, NWT (Figure 1.1). They focussed on controlling factors for this ice enrichment such as soil texture, soil moisture status, ground temperature, and organic matter content.

The top of permafrost may be enriched with segregated ice by two processes. First, ice lenses frozen at the base of the active layer may be trapped when the permafrost table rises, forming aggradational ice (Mackay, 1972). Second, a moisture imbalance caused by unequal downward (summer), and upward (winter) unfrozen water migration under opposing temperature gradients may lead to gradual ice enrichment at the top of permafrost by repeated segregation (Cheng, 1983). The permafrost table may rise with a decrease in active-layer thickness, or when the ground surface aggrades from sediment deposition or vegetation growth (Cheng, 1983; Mackay and Burn, 2002a; Kokelj and Burn, 2005). Over time, a rising permafrost table and repeated ice segregation may result in the development of a thick ice-rich zone at the top of perennially frozen ground (Cheng, 1983; Burn, 1988).

The ice-rich zone is of practical interest because its thaw can lead to extensive hillside failure and thermokarst subsidence (Mackay, 1970, 1972; Haeberli and Burn,

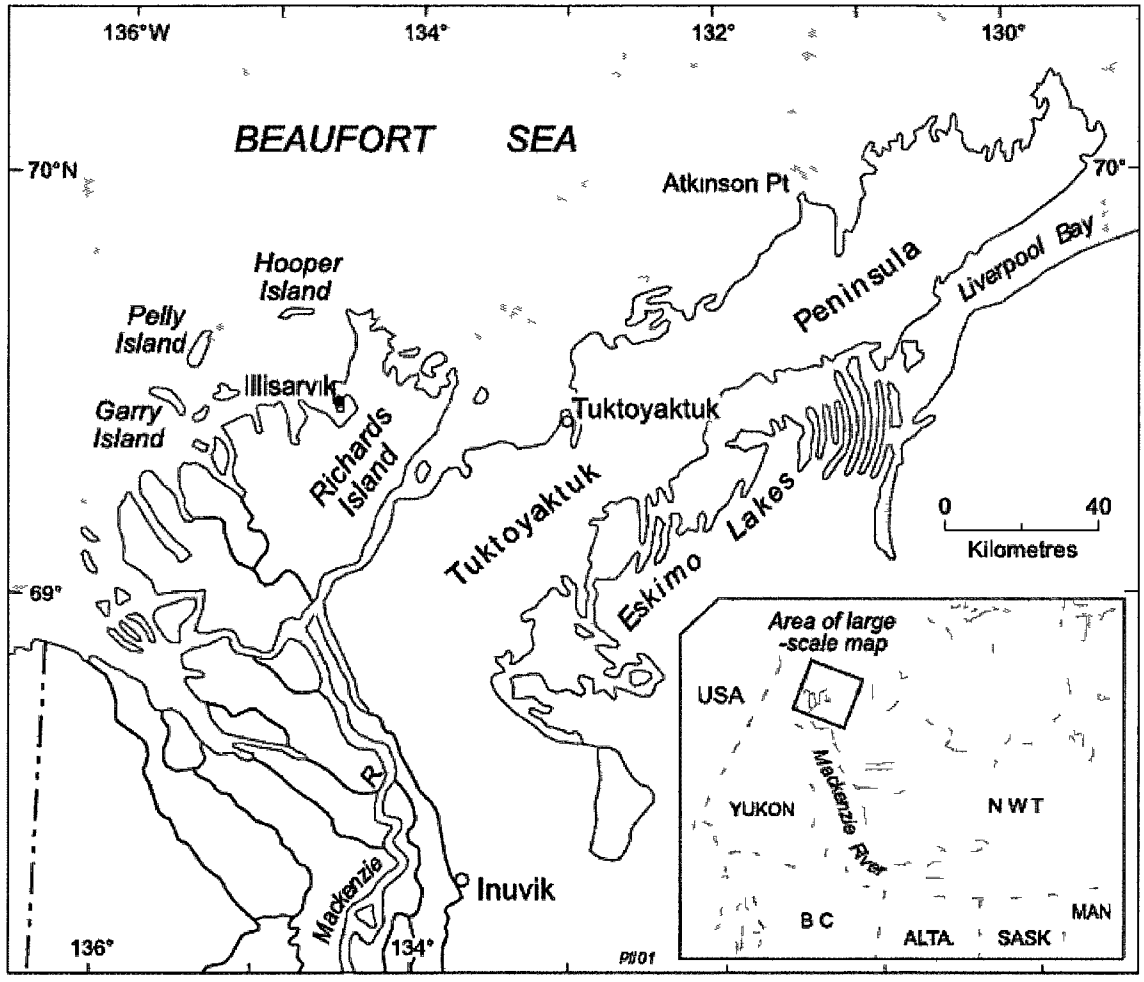


Figure 1.1 Location of study area (from Mackay and Burn, 2002b).

2002). This may damage infrastructure such as roads, pipelines and buildings. Considering the rising temperatures in the Arctic and the increasing number of northern development projects, such as the proposed Mackenzie Gas Project and the extension of the Dempster Highway to Tuktoyaktuk, better understanding of near-surface ground ice is necessary for planning purposes and hazard mitigation.

Near-surface ground ice also influences the ecology, morphology and hydrology of the landscape. Melted ice may supply the active layer with nutrients and water, and promote shrub growth (Mackay, 1995; Sugimoto et al., 2002: Kokelj and Burn, 2003). A positive feedback that enhances permafrost degradation and ecological succession may occur as larger shrubs trap more snow, further warming the ground and accelerating their growth (Sturm et al., 2001).

1.2 Background research

Aggradational ice in permafrost was first described by Mackay (1972). Later, Cheng (1983) and Mackay (1983) provided evidence of net seasonal migration of moisture downward from the active layer causing gradual ice enrichment of near-surface permafrost. Frost heave associated with the migration of liquid moisture and growth of ice in near-surface permafrost was observed in several investigations during the 1980s (e.g. Mackay, 1983; Smith, 1985; Burn, 1988). More recently, interest has focused on the rates of ice accumulation and the effect of ice growth on vegetation structure and earth hummock morphology (Kokelj and Burn, 2003a; Kokelj et al., 2007).

The ice-rich zone at the top of permafrost has been classified by Shur et al. (2005) as the *transition zone*, situated between the active layer and underlying long-term

permafrost. The base of this zone is marked by the top of primary ice wedges or other massive ice bodies. The upper and lower bounds of the transition zone may also be determined from botanical and geochemical discontinuities (Burn, 1988, 1997; Kokelj and Burn, 2003). In the transition zone classification, the top of permafrost acts as a thermal buffer to protect deeper permafrost from thaw, due to the latent heat requirement of ice-rich ground (Shur et al., 2005). The uppermost part of the transition zone, termed the *transient layer*, has a lower ice content than the rest of the transition zone because it undergoes thaw more frequently (Shur et al., 2005).

There are only a few estimates of the rate of ice accumulation in permafrost under natural conditions. Burn (1988) estimated that 0.1 to 0.2 mm yr^{-1} of water had been incorporated into permafrost over the past 7-8000 years in glaciolacustrine sediments near Mayo, YT, and Mackay and Burn (2002a) reported 5 mm yr^{-1} of ice growth over twenty years in the Illisarvik experimentally drained lake from 1978 to 1998. Kokelj and Burn (2003) estimated the rate of water incorporation to permafrost in hummocky terrain near Inuvik, NWT. The excess ice, which is the ground-ice content in excess of the saturated moisture content of thawed soil, present above a 1981 thaw unconformity indicated that water incorporation was between 0.5 and 2 mm yr^{-1} over two decades. The relative importance of factors controlling the rate of accumulation has not been examined.

1.3 The Illisarvik drained lake experiment

The field work for this research took place at Illisarvik, the experimentally drained lake basin on Richards Island, near the western Arctic coast (Figure 1.2). The

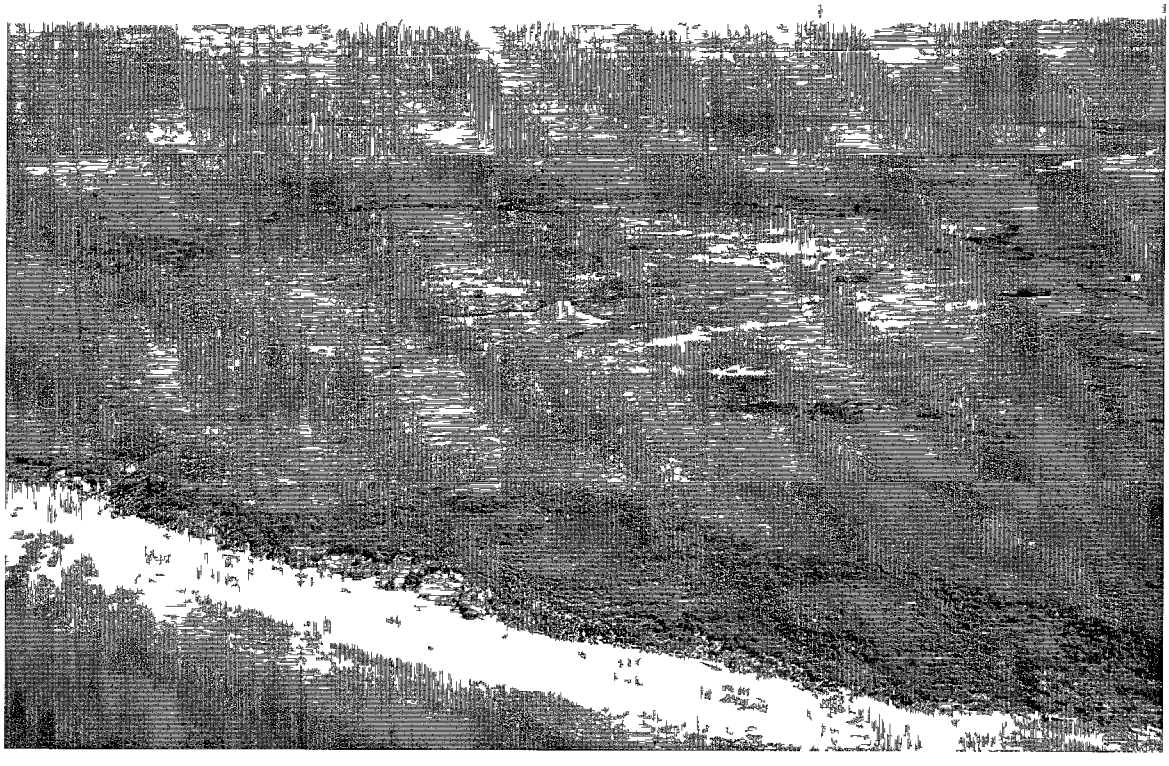


Figure 1.2 The Illısarvik drained lake basin in 2008. Photo: C.R. Burn.

lake was drained by Dr. J. Ross Mackay on August 13, 1978, in order to study the growth of permafrost and related phenomena (Mackay, 1997). Prior to drainage, the lake was underlain by a bowl-shaped talik, 32 m deep near the centre of the lake basin (Mackay, 1997). The lake basin therefore offered a unique opportunity to study the growth of ground ice in a temporally controlled setting, as near-surface ice present in the basin sediments has formed since the drainage of the lake. In addition, active-layer depths have been recorded regularly along the basin axes since drainage (Mackay and Burn, 2002a). The relations between active-layer depths and the development of near-surface ground ice can consequently be examined.

1.4 Research objectives

The purpose of this research is to investigate the factors that control near-surface ice enrichment at the Illisarvik site, and to study the development of near-surface ice over the past 32 years. The ground-ice characteristics of near-surface permafrost were sampled by coring at 38 sites in the drained lake and surrounding landscape.

The specific research objectives of this thesis are to: (1) examine the relations between soil texture, wetness, ground temperature, and organic matter content in the lake bed and near-surface ground ice content; (2) determine the rate of near-surface ice accumulation in permafrost over the past 32 years in the lake basin and surrounding tundra; and (3) compare ground ice conditions in the lake basin to those in the surrounding tundra. Since the development of near-surface ground ice is inherently related to active-layer dynamics, an examination of the active-layer history in the Illisarvik lake basin is included in this study.

1.5 Thesis structure

The thesis is comprised of six chapters. The following chapter presents a review of the processes of near-surface ground ice formation, and their relations with terrain morphology, ecology, and hydrology. Chapter 3 describes the study area and the study design. Chapter 4 discusses the basin active-layer history, the present ice conditions at the sampling sites, the relations between controlling factors on ice accumulation, and the growth of ice over thirty years. Chapter 5 is a discussion of the study results, and their implications in the context of near-surface ground ice studies. Finally, Chapter 6 provides of a summary of the results, and presents conclusions and suggestions for further research.

2. NEAR-SURFACE GROUND ICE

2.1 Introduction

A study of the development of near-surface ground ice requires understanding of: (1) permafrost characteristics and classifications; (2) the nature of moisture movement in frozen soil and processes responsible for segregated-ice formation; (3) Darcy's Law, which dictates several controls on ice accumulation; and (4) the relations between near-surface ground ice, terrain morphology, ecology and hydrology. This chapter reviews the conceptual models of permafrost terrain, the migration of unfrozen water and ice segregation, the controlling factors of ice development, and the influence of near-surface ground ice on landscape form and function.

2.2 Permafrost and the active layer

2.2.1 Two-layer conceptual model

The traditional conceptual model of permafrost terrain employs a two-layer, quantitative classification based on temperature (Figure 2.1). Permafrost is defined as ground that remains at or below 0°C for at least two consecutive years (ACGR, 1988). The ground above permafrost that freezes and thaws annually is the active layer, and its thickness varies with air temperature, vegetation, snow cover, substrate type, slope, aspect, and soil water content (Mackay, 1982; French, 2007). The active-layer begins freezing downward in autumn, but may also freeze upward from the top of permafrost (Mackay, 1984). Upward freezing is common in cold permafrost, such as the western Arctic coast, but not as important in warm permafrost, such as in the discontinuous

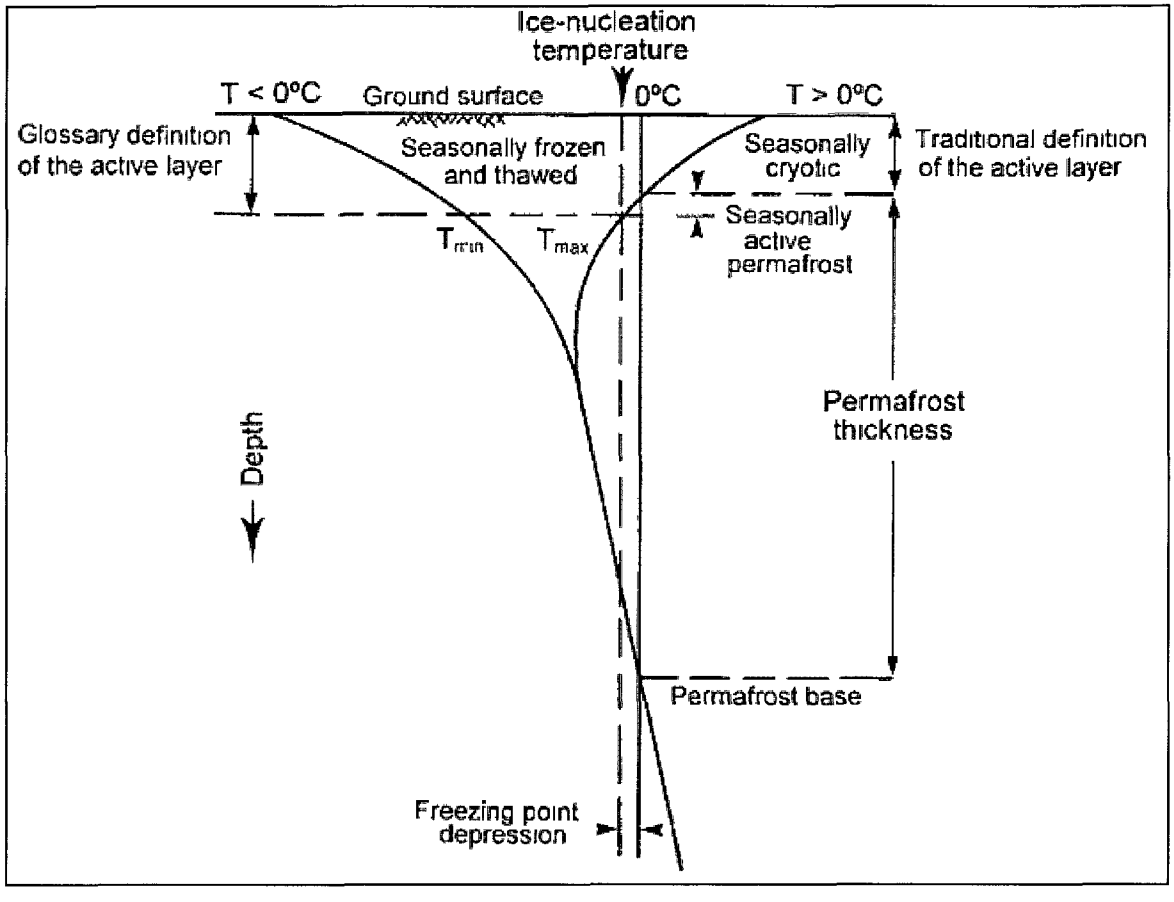


Figure 2.1 Terms used to describe ground temperatures and states of water in permafrost (after ACGR, 1988, Figure 2; and Burn, 1998, Figure 1).

permafrost zone. Segregated ice lenses form near the ground surface from downward freezing, and at the base of the active layer if upward freezing occurs (Mackay, 1983).

2.2.2 The transition zone conceptual model

Shur et al. (2005) presented a three-layer, qualitative conceptual model for the surface of permafrost terrain, which includes the active layer, the *transition zone*, and long-term permafrost (Figure 2.2). This classification is based on the frequency of active-layer depths and ice content of the upper permafrost. Below the active layer, the top layer of permafrost is termed the transition zone. The lower boundary of this layer is coincident with the long-term permafrost table, marked by tops of primary ice wedges (Shur et al., 2005). The transition zone is ice-rich, containing segregated ice lenses, ice veins, and contemporary ice wedges. Thaw of the transition zone occurs at sub-decadal to multi-centennial timescales (Shur et al., 2005).

Shur et al. (2005) define the top of the transition zone as the *transient layer* after Yanovsky (1933), where thaw is more frequent than in deeper parts of the transition zone. As a result, Shur et al. (2005) propose that the transient layer is less ice-rich than the rest of the transition zone. Deep thaw in the transition zone is largely controlled by climatic changes at long timescales (Shur et al., 2005). The transient layer, situated immediately below the active layer, protects the deeper parts of the transition zone from thaw during most warm years (Shur et al., 2005).

2.2.3 Climate and ecosystem-driven permafrost

Shur and Jorgenson (2007) created a classification system to describe the interactions of climate and ecology as they relate to the formation and degradation of

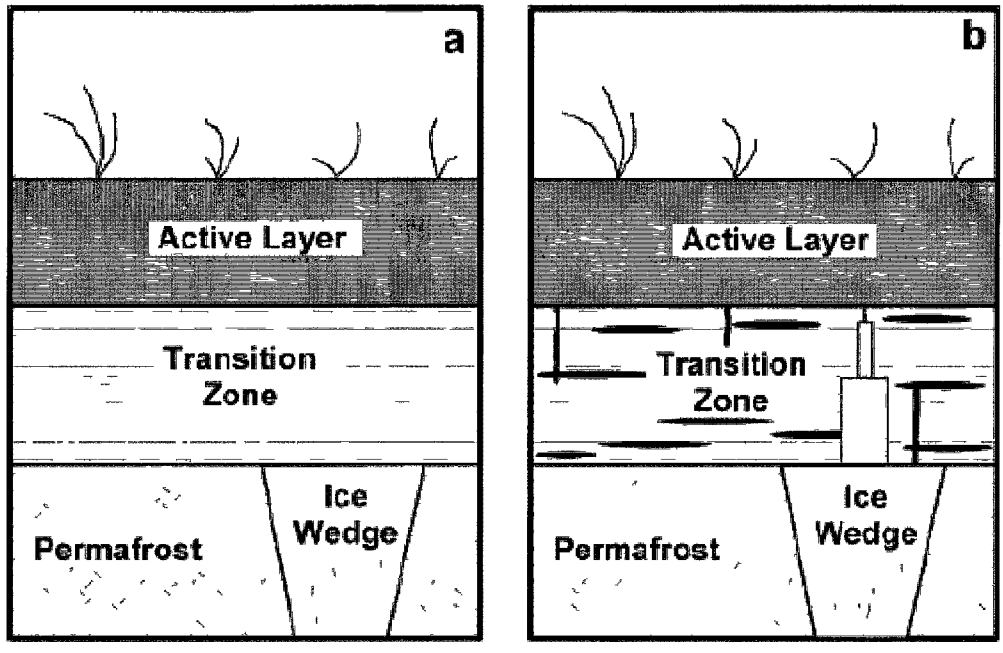


Figure 2.2 Three-layer conceptual model showing the transition zone (a) immediately following deep thaw and (b) after ice enrichment over many centuries (after Shur et al., 2005). Black ovals represent ice lenses and black vertical lines represent ice veins.

permafrost. The classification scheme identifies five types of permafrost, including *climate-driven permafrost,* and *climate-driven, ecosystem-modified permafrost.*

Climate-driven permafrost forms in the continuous permafrost zone immediately following exposure of an unfrozen soil surface (Shur and Jorgenson, 2007). This type of permafrost exists in the High Arctic, and initially forms on barren surfaces in the Low Arctic before ecological succession occurs (Shur and Jorgenson, 2007). Climate-driven permafrost is sensitive to rapid climatic change if warming occurs faster than vegetation succession. In contrast, it is insensitive to surface disturbances due to the absence of an insulating organic or vegetation layer. The potential for thaw subsidence is low because the permafrost table has not aggraded significantly, limiting aggradational ice formation (Shur and Jorgenson, 2007).

Climate-driven, ecosystem-modified permafrost is present in the continuous permafrost zone, when vegetation growth and organic accumulation have modified the surface (Shur and Jorgenson, 2007). The vegetation succession is normally accompanied by the formation of aggradational ice if the active layer thins. The active layer may thin from the insulating effect of vegetation and surface organic matter (e.g. Price, 1971), or from a cooling climate. Therefore, climate-driven, ecosystem modified permafrost is sensitive to surface disturbances such as fire or vegetation removal, and increasing temperatures. These changes would all lead to thaw subsidence as the active layer thickens and near-surface ice melts (Mackay, 1970).

2.3 The formation of segregated ice

2.3.1 Unfrozen water in cryotic ground

When the ground temperature falls below 0°C, some water in the soil remains liquid in small pores, and as layers at the edge of soil particles. Solutes in ground water lower the freezing point, and restrictions on nucleation within small soil pores prevent ice formation at 0°C. Most importantly, electrochemical forces emanating from soil particles cause films of unfrozen water to persist at sub-zero temperatures (Mitchell and Soga, 2005). Solutes are expelled from the ice phase as crystals form in soil pores, resulting in a freezing-point depression in the remaining water. The effect of these solutes is generally minimal and the freezing point is only depressed about 0.1°C (Williams and Smith, 1989; French, 2007).

The restricted volume of soil pores prevents ice nucleation at 0°C as a result of capillary forces. More important, however, are electrochemical forces exerted by soil particles, which hold water in films under tension, and lower the freezing point of pore water (French, 2007). This adsorbed water persists as films on soil particles as a result of four mechanisms (Mitchell and Soga, 2005). First, soil particle surfaces possess unbalanced charges, and positively charged sides of bipolar water molecules are attracted to soil particles (Figure 2.3a). This attraction prevents the hydrogen bonding necessary for the formation of an ice lattice. Second, hydration occurs between the mineral molecules and pore-water solutes, limiting ice formation (Figure 2.3b). Third, cations in pore water are attracted to the negatively-charged soil particle surfaces, further reducing the freezing point (Figure 2.3c). Finally, dipolar attraction is responsible for some water held to the negatively charged clay particles (Figure 2.3d). Adsorbed cations are tightly

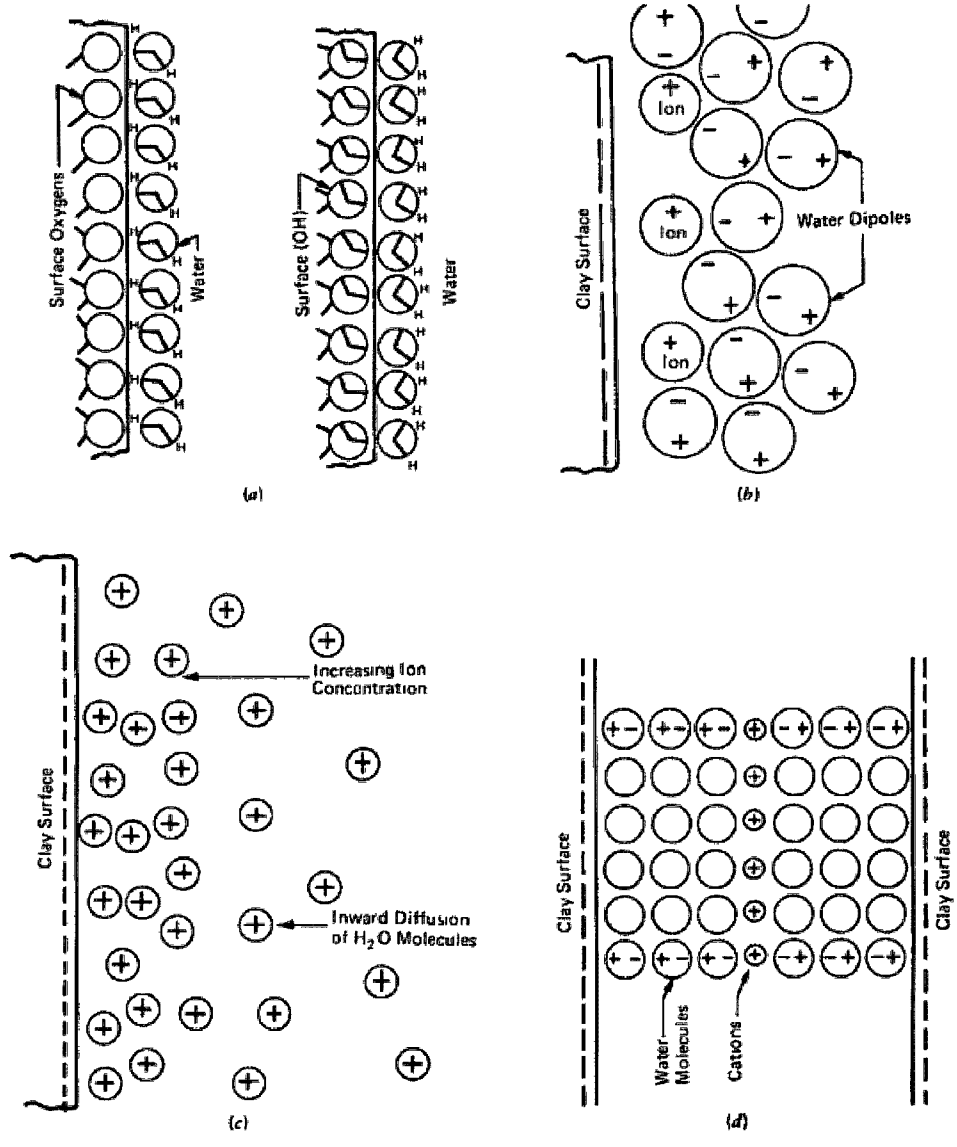


Figure 2.3 Mechanisms of water adsorption by clay minerals (a) hydrogen bonding, (b) ion hydration, (c) attraction by osmosis, and (d) dipole attraction (Mitchell and Soga, 2005, Figure 6.4).

held against the surface of negatively charged clay particles. These particles try to diffuse away into the pore water, however they are restricted by their attraction to the negatively charged surface particles (Mitchell and Soga, 2005). The result is an increase in cations with proximity to the soil particle surface and an associated decrease in freezing point. The charged soil surface and distributed charge in the pore water are termed the *diffuse double layer* (Mitchell and Soga, 2005).

The specific surface area (SSA) – the surface area of particles per unit volume of soil – is the primary factor controlling the amount of adsorptive forces in the soil. Clay, having a greater SSA than sand or silt, holds more adsorbed water under higher tension and therefore contains the most unfrozen water at temperatures below 0°C (Figure 2.4). Silt also holds an appreciable amount of unfrozen water at temperatures below 0°C, though it is less than clay because of a lower SSA (Figure 2.4). Because sand has such a low SSA, it holds little unfrozen water below 0°C (Williams and Smith, 1989). As the temperature falls and soil water is turned to ice, adsorbed films become thinner and the unfrozen water content decreases (Williams and Smith, 1989).

2.3.2 Water migration in cryotic soil

The flow of fluid through a porous medium, such as the flow of water in frozen soil, is governed by Darcy's law:

(1) $$Q = AK\frac{\Delta h}{L}$$

where Q is the volumetric flow rate in $m^3\ s^{-1}$, A is the cross-sectional area perpendicular to flow, K is the hydraulic conductivity, Δh is the hydraulic head, and L is the flow path length. The fraction $\Delta h\ L^{-1}$ is the pressure gradient.

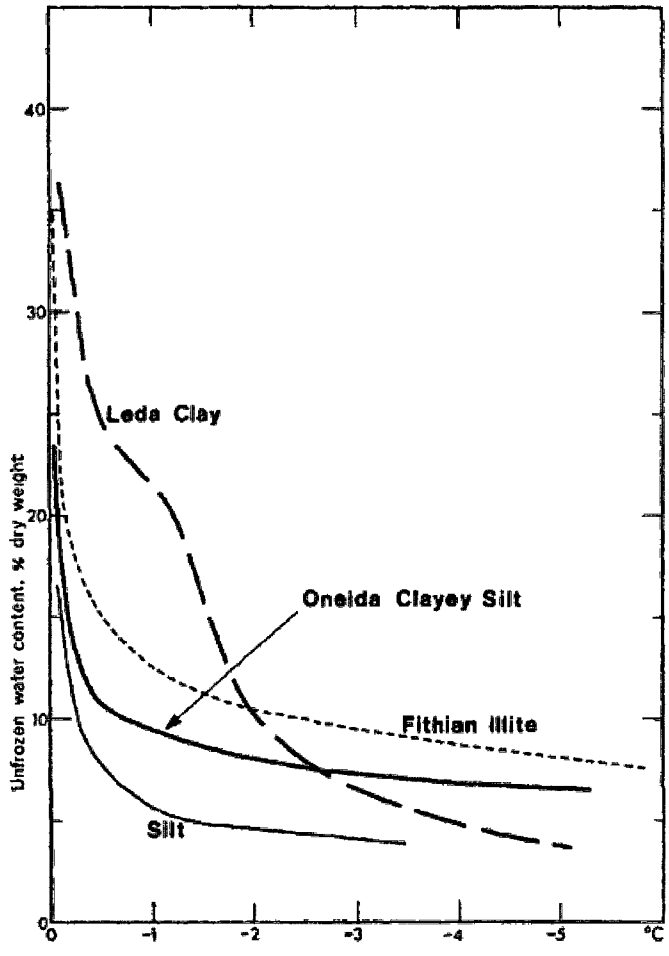


Figure 2.4 Unfrozen water content curves for different soil types (Burt and Williams, 1976, Figure 1).

The hydraulic conductivity of frozen soil is controlled by soil type and temperature (Figure 2.5). Soil type influences the hydraulic conductivity because it governs the unfrozen water content below 0°C, and the potential for groundwater movement. However, the structure of the soil is also important to hydraulic conductivity. For example, clay has greater unfrozen water content than silt or sand below 0°C, but it has a lower hydraulic conductivity than silt because the pore structure is more tortuous (Burt and Williams, 1976) (Figure 2.5). Silts have the highest frozen hydraulic conductivities because they possess relatively high unfrozen water contents and are more permeable due to greater pore space and connectivity (Burt and Williams, 1976).

The pressure gradient in frozen soil is principally a result of capillary and adsorptive forces, which hold the unfrozen water under tension (suction). In the field, the pressure gradient is induced by the thermal gradient in the ground (Cheng, 1983). At higher temperatures, the adsorbed water films are thicker and under higher pressure (lower tension) than thinner films at lower temperatures. As a result, unfrozen water migrates along the temperature-induced pressure gradient in the direction of decreasing temperature (Cheng, 1983; Mackay, 1983). The change in pressure of unfrozen water with temperature is often given as 1.2 MPa $°C^{-1}$ (Edelfsen and Anderson, 1943).

2.3.3 Ice segregation

The migration of unfrozen water to lower temperatures forms segregated ice lenses. Ice lenses form at temperatures below 0°C, and the majority of water transport between the 0°C isotherm and the first (warmest) ice lens occurs in the liquid phase (Burn, 1986). This zone comprises two parts, the first between the 0°C isotherm and the

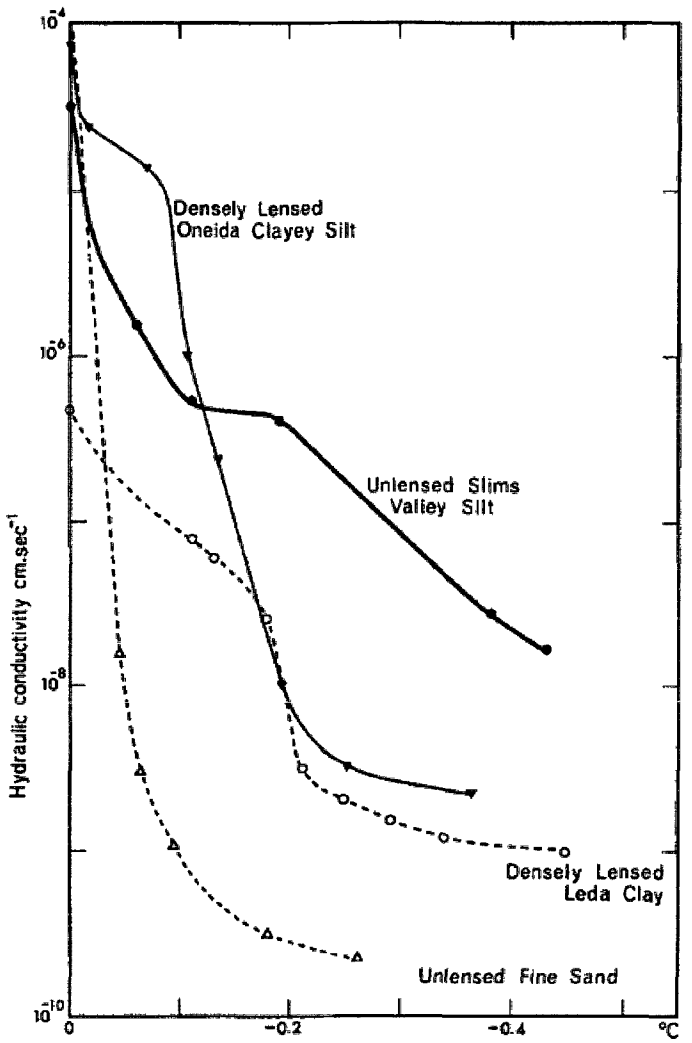


Figure 2.5 Soil type and hydraulic conductivity of frozen soils (Burt and Williams, 1976, Figure 6). In silty clay near Inuvik, NWT, hydraulic conductivity was found to decline to 3.5×10^{-12} m/s at -1°C by Smith, 1985.

warmest pore ice, and the second between the pore ice and the warm side of the ice lens, termed the *frozen fringe* (Figure 2.6) (O'Neill and Miller, 1985). Liquid water transport towards the ice lens occurs in the frozen fringe in unfrozen pore water or in adsorbed water films around particles (Cheng, 1983; O'Neill and Miller, 1985).

The thickness of growing ice lenses depends on the speed of the 0°C isotherm advance through the soil. The frozen fringe thickens as the frost front advances and the ice lens grows at progressively lower temperatures. The hydraulic conductivity is lower at lower temperatures, and consequently the growth of the lens slows. Following this, the location of lens growth shifts closer to the 0°C isotherm, where the hydraulic conductivity is greater, and a new lens forms under more favourable thermal, hydrological and mechanical conditions near the freezing front (Miller, 1972; 1978). Therefore, when the frost front penetrates the soil slowly, more time is available to grow each ice lens before a new one forms.

Growth of segregated ice results in upward heave of the ground surface. Migration of water in frozen soil and subsequent frost heave from segregated-ice formation has been reported in several field studies (Cheng, 1983; Mackay, 1983; Smith, 1985).

2.4 Mechanisms of segregated ice accumulation at the top of permafrost

Segregated ice may accumulate at the top of permafrost as a result of two distinct processes. Ice enrichment may occur: (1) each year from a seasonal moisture imbalance supplying the upper permafrost, forming *repeated-segregation ice* (Cheng, 1983), or

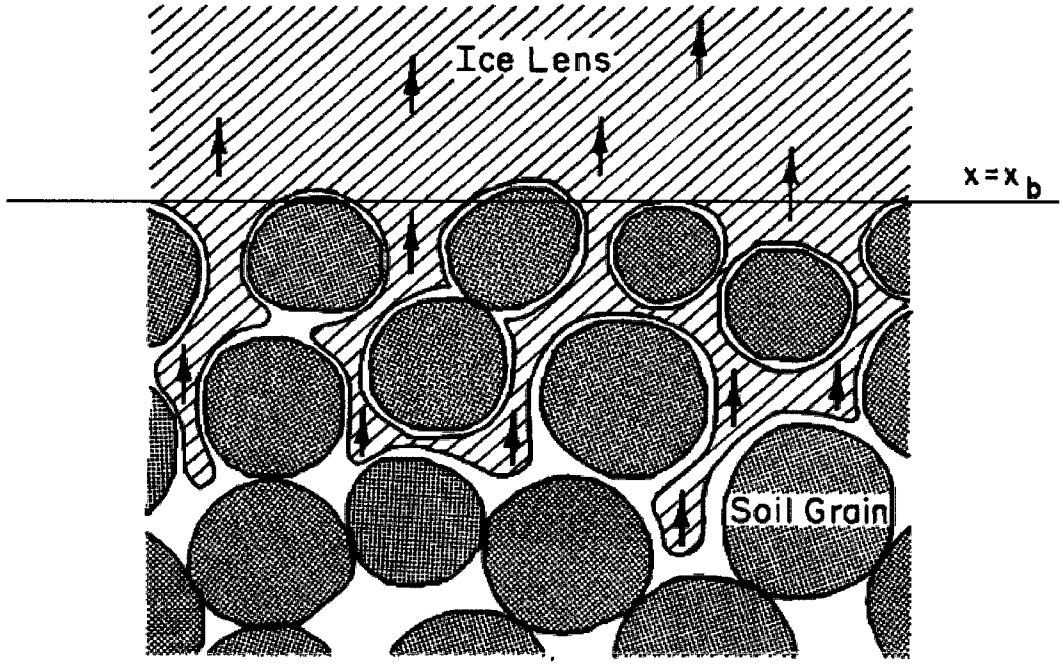


Figure 2.6 Diagram of the frozen fringe. $X = X_b$ indicates the position of the lens-frozen soil interface. The arrows represent the movement of ice by regelation (O'Neill and Miller, 1985, Figure 1).

(2) from the trapping of ice lenses when the permafrost table rises, forming *aggradational ice* (Mackay, 1972). The combination of repeated-segregation ice and aggradational ice growth can lead to thick layers of ice-rich ground (Cheng, 1983).

2.4.1 Repeated-segregation ice

Repeated segregation ice accumulates from a net annual migration of moisture to the top of permafrost (Cheng, 1983). Water may be drawn into the still-frozen active layer during the snowmelt season by temperature-induced suction gradients (Burn and Smith, 1985). In the summer, these gradients draw water from the thawing active layer to the frozen ground below, ahead of the thawing front (Cheng, 1983). The opposite occurs in winter when the active layer is frozen; moisture is transported upward in adsorbed water films, from the permafrost into the active layer along the ground-temperature gradient (Parmuzina, 1978; Mackay, 1983). The upward migration in winter is limited because the hydraulic conductivity is significantly lower at winter temperatures (Burt and Williams, 1976; Cheng, 1983). In addition, moisture from snowmelt, precipitation and ice lens thaw may supply downward migration in summer, whereas a hydrologically closed system exists in winter, and the supply of moisture is limited to water already in the ground (Cheng, 1983). The key time for moisture incorporation to the top of permafrost is in late summer or early fall, when the thaw depth is near maximum and the base of the active layer is commonly saturated from ice lens thaw and precipitation. As a result of this unequal migration, the top of permafrost may be enriched with ice over time.

2.4.2 Aggradational ice

Aggradational ice forms when the permafrost table rises and traps segregated ice lenses formed at the base of the active layer during upward freezing (Figure 2.7). The formation of aggradational ice is therefore linked to climate cooling, the growth of an insulating organic layer, or the accumulation of sediment or other material at the ground surface (Mackay, 1972).

2.4.3 Rates of ice accumulation

Monitoring of ice enrichment at the top of permafrost has been reported by Mackay and Burn (2002a), who observed the ice content at several points over twenty years at Illisarvik. Following drainage in 1978, the ice content at the top of permafrost increased until 1993, where it remained steady until a slight decline in 1997 and 1998 due to deep summer thaw depths. The rate of accumulation over the study period was estimated at 5 mm y^{-1} (Mackay and Burn, 2002a). Recently, Kokelj and Burn (2003) estimated the rate of water incorporation to permafrost in hummocky terrain near Inuvik, NWT. The excess ice present above a 1981 thaw unconformity indicated that water incorporation was between 0.5 and 2 mm yr^{-1} over two decades. The vertical extent of ice accumulation was limited to 10 cm at the top of permafrost, which likely formed between 1991 and 2001 when the active layer was static. Permafrost was ice-poor in the 20 cm above the thaw unconformity where the active layer thinned quickly (Kokelj and Burn, 2003).

A long-term rate of ice growth since the early-Holocene climatic optimum around 8000 years ago was estimated by Burn (1988) at an exposed thaw-slump headwall near

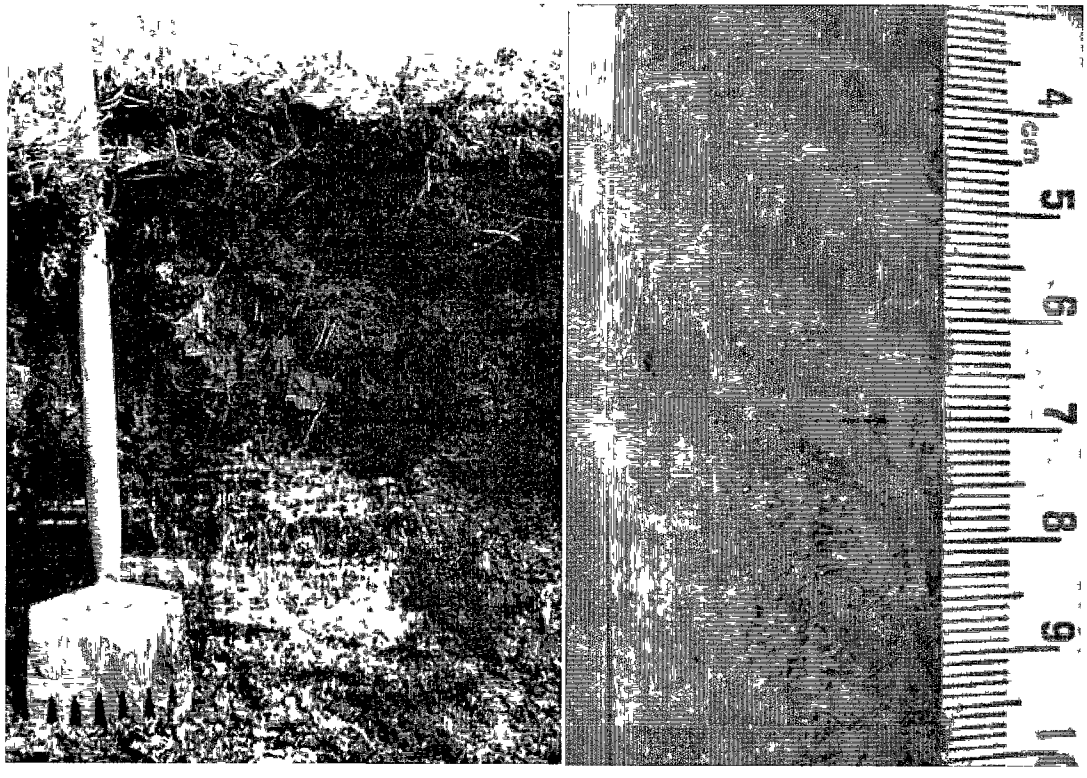


Figure 2.7 Aggradational ice (left), visible as white bands, immediately below the permafrost-active layer boundary (Mackay, 1972, Figure 10). Near-surface ground ice lenses (right), visible as the darker areas. The top of permafrost is located at approximately 7 cm on the ruler (Morse et al., 2009, Figure 2).

Mayo, YT. The amount of ice in the headwall above a thaw unconformity corresponded to 0.1 to 0.2 mm of water added per year in the past 7 or 8 thousand years (Burn, 1988).

Burn (1988) also estimated that approximately 10 mm yr^{-1} of moisture was incorporated into permafrost over two years in soil tubes filled with a silt-loam slurry. The results of the soil tube experiment are likely indicative of the maximum possible water migration because moisture was not limited, and the soil used was particularly frost-susceptible (Burn, 1988).

The range of moisture accumulation rates reported over different timescales suggest that the initial rate may not be indicative of long-term conditions (Kokelj and Burn, 2003). The soil-tube estimate is an order of magnitude larger than the rates measured at Illisarvik (Mackay and Burn, 2002a) and near Inuvik (Kokelj and Burn, 2003) at decadal timescales. In addition, the rate of accumulation observed in glaciolacustrine sediments near Mayo (Burn, 1988) over the millennial timescale is up to an order of magnitude less than the decadal rates.

2.5 Controlling factors of ice accumulation

2.5.1 Soil texture

Soil type is an important control on segregated ice formation due to the influence of particle size on the unfrozen water content, which controls in part the hydraulic conductivity (Figure 2.5). Fine grained soils have a greater particle specific surface area than coarse-grained material, and hold more adsorbed water (Williams and Smith, 1989). Therefore, fine-grained soil has more free water available for migration towards the

freezing front at temperatures below 0°C. Hydraulic conductivity is greatest in silts, so water can more readily reach growing ice lenses, making silty soil the most frost-susceptible.

Results from near-surface ground ice field studies have found relations between soil texture and near-surface ground-ice content. Sediment texture influenced the excess-ice content at the top of permafrost at over 70 sites in the Mackenzie River delta (Kokelj and Burn, 2005). The excess-ice content decreased as the proportion of sand increased, and silt was associated with higher excess-ice contents. However, ice-rich sediments were associated with a range of silt and clay contents, which indicates that other factors are also important controls on near-surface ground ice (Kokelj and Burn, 2005). Similarly, Morse et al. (2009) found that ice-rich permafrost in the outer Mackenzie Delta was associated with soils usually composed of at least 40% silt, and in the Illisarvik drained lake basin, aggradational ice has formed mainly in frost-susceptible organic-rich silts, with little found in sandy areas (Mackay and Burn, 2002a).

2.5.2 Temperature

As temperatures decrease below 0°C, the adsorbed water layers on soil particles become thinner, and the unfrozen water content and hydraulic conductivity decline rapidly (Burt and Williams, 1976; Smith, 1985). The apparent hydraulic conductivity of frozen silty clay was estimated from field data by Smith (1985) at Inuvik, NWT. The calculation of hydraulic conductivity was based on Darcy's Law, and determined from the rate of water migration, obtained from soil heave data, and the water head, calculated from the mean temperature gradient. The hydraulic conductivity was found to decline from 7×10^{-9} m/s just above freezing to 3.5×10^{-12} m/s at -1°C (Smith, 1985).

The seasonal difference in frozen hydraulic conductivity is important to repeated-segregation ice formation because upward migration of unfrozen water away from the top of permafrost may occur in winter in permafrost near 0°C. This upward migration is limited under colder winter conditions (Cheng, 1983). Therefore, it is expected that areas underlain by permafrost near 0°C will have less repeated-segregation ice than colder permafrost due to the greater winter upward migration of water.

Kokelj et al. (2005) examined the ground-ice content in willow/alder communities on aggrading point bars, and in spruce forests and lakeside alder areas in the Mackenzie delta. The willow and alder communities were underlain by 'warm' permafrost, where the annual temperature was greater than -2°C (Kokelj and Burn, 2005). Colder permafrost, which was well below -2°C, existed below spruce forests and lakeside alder sample areas (Kokelj and Burn, 2005; Fig. 8). The decrease in unfrozen water content at the top of permafrost between summer and winter was fivefold in areas with cold permafrost (Kokelj and Burn, 2005). In the areas with warmer permafrost, the unfrozen water content change was only 5% (Kokelj and Burn, 2005). Little excess-ice was observed at sites in willow and alder areas, and ice contents were generally less than 20%. Higher excess-ice contents, up to 60%, were observed at spruce forests and lakeside alder sites.

At Illisarvik, the growth of vegetation in some areas of the basin has resulted in snow entrapment and the warming of permafrost. Mackay and Burn (2002) presented permafrost temperatures from four thermistor cables located in presently well vegetated ground in the north end of the basin. The four thermistor cables were arranged in a row and increased in distance from the lake shore. In 1980, permafrost temperatures from 6-8

m depth were -3°C to -4°C at the two sites nearest to the lake shore. By 1985-86, permafrost temperatures at these locations had cooled to below -5°C and -6°C as a result of permafrost aggradation (Mackay and Burn, 2002). Subsequently, the permafrost temperatures rose to above -3°C and -4°C by 1999 (Mackay and Burn, 2002; Fig. 18). At the two sites further from the lake shore, permafrost temperatures from 6-8 m depth decreased to around -3°C in the early 1990s and then increased to above -2.5°C in 1995 to 1998 (Mackay and Burn, 2002). The warming began first at the sites nearest to the lake shore, and progressively moved away from the lakeshore. This trend was concurrent with the growth of vegetation and snow entrapment progressively outwards from the lakeshore. Conversely, in the basin centre, where vegetation growth and snow entrapment is limited, permafrost has continued to cool in the past decade. At one basin centre site, the permafrost at 5 m depth cooled from approximately -1.5°C to -2.7°C between August 2000 and August 2005 as a result of permafrost aggradation (Parameswaran and Burn, 2010).

The observed warming of near-surface permafrost from vegetation establishment and snow entrapment in some areas of the Illisarvik lake basin may therefore allow greater upward water migration away from the top of permafrost, limiting the accumulation of repeated-segregation ice compared to areas with little vegetation and colder permafrost.

It should be noted that ground temperatures may be indicative of the age of permafrost. Kanigan et al. (2009) observed higher permafrost temperatures at sites in the Mackenzie delta with newer permafrost and lower temperatures where older permafrost existed. Higher temperatures in younger permafrost were due to deep snow cover and the

latent heat requirement to freeze thick active layers (Kanigan et al., 2009). In such cases, the accumulation of repeated-segregation ice may be limited not only by the upward migration of water in the winter, but also by the limited number of annual cycles transporting moisture to the top of permafrost. When comparing ice accumulation between sites with warm and cold permafrost, the effect on unfrozen water contents *and* the potential age differences of the permafrost must therefore be considered. At Illisarvik, all basin permafrost has aggraded since 1978, and is therefore of equal age.

2.5.3 Soil moisture status

Soil moisture characteristics have been found to greatly influence near-surface ground ice content. Moisture addition to the top of permafrost can be supplied by ice lens thaw in the active layer, snow melt, rainfall, or groundwater flow (Cheng, 1983). Though some water may be transported upwards during winter from permafrost at depth, the majority of the water involved in ice segregation originates in the active layer, from where it more readily migrates downward in summer (Cheng, 1983).

Kokelj and Burn (2005) concluded that a thick layer of aggradational ice in a lakeshore/wetland area resulted from upward permafrost aggradation in a saturated environment. Simulations of near-surface ground ice growth under field conditions have used PVC tubes filled with a silt-loam slurry, set into the ground (Burn, 1988). The soil moisture in the tubes was initially above saturation, and ice lenses developed at the top of permafrost at rates around 10 mm yr^{-1} (Burn, 1988). This idealized simulation indicates that ice growth is rapid when water is abundant. These studies indicate that the availability of unfrozen water in the active layer is an important control on ice formation.

2.5.4 Active-layer history

The formation of aggradational ice necessitates a rise in the permafrost table to trap ice lenses at the base of the active layer (Mackay, 1972). Therefore, active-layer history is a control on aggradational ice formation.

The importance of active-layer history over a long timescale is demonstrated by Burn (1988), who observed 90 cm of segregated ice above an early Holocene thaw unconformity in the headwall of a thaw-slump near Mayo, YT. The ice likely accumulated since the early-Holocene climatic optimum, around 8000 years ago, as the active-layer thinned under the gradually cooling climate that followed (Burn, 1988).

Kokelj and Burn (2003) observed the importance of active-layer history to ice accumulation above a 1981 thaw unconformity in permafrost near Inuvik, NWT. Immediately above the thaw unconformity, permafrost was ice-poor where the active layer had thinned rapidly. Above the ice-poor zone, excess-ice had accumulated where the active layer was stable between 1991 and 2001 (Kokelj and Burn, 2003). These observations indicate that rapid aggradation of the permafrost table limits ice accumulation, which was also suggested by Cheng (1983)

Over a short time period, near-surface ground ice forming in the Illisarvik drained lake basin between 1979 and 1999 resulted in the uplift of the ground surface by up to 10-15 cm (Mackay and Burn, 2002a). Abnormally high temperatures in 1998, due to a strong El Niño event, resulted in deep active layers in the Mackenzie Delta area that year (Smith et al., 2001). The increased thaw penetration melted some of the newly accumulated near-surface ground ice at Illisarvik and caused ground subsidence (Mackay

and Burn, 2002a). Further discussion regarding active layers at Illisarvik follows in the next three chapters.

These examples illustrate how variation in active-layer depths and long-term active-layer history may influence the accumulation and thaw of near-surface ground ice.

2.5.5 Organic matter content

Soil organic matter is characterized by low bulk density and high porosity compared to mineral particles. These characteristics have implications for the measurement and interpretation of ground-ice data. First, variation in gravimetric moisture content is related to organic matter content due to its low bulk density (Morse et al., 2009). Gravimetric moisture content data must be interpreted carefully because high moisture content may be indicative of high organic matter content rather than high ice content.

The high porosity of organic matter may also influence the measured excess-ice content. Excess ice is commonly calculated by measuring the volume of supernatant water that results after a sample of permafrost has melted (e.g. Mackay and Burn, 2002; Kokelj and Burn, 2005). However, soil water retention is correlated with organic matter content (Gupta and Larsen, 1979). Consequently, organic-rich samples of permafrost may yield little or no supernatant water when thawed compared to less organic samples, even though the volumetric ice content may be high. The low bulk density and water retention capability of organic matter must therefore be considered when examining gravimetric moisture and excess-ice content data in organic-rich soils.

2.6 Physical consequences of near-surface ground ice

The accumulation of excess-ice in near-surface permafrost causes uplift of the ground surface. Heave and subsidence from excess-ice growth or thaw, respectively, may be problematic to infrastructure such as roads, houses, and pipelines. The accumulation of excess ice occurs slowly, and short-term implications may be small. In contrast, the thawing of near-surface ground ice may occur more rapidly over a warm summer or several warm years, and the resulting subsidence may pose serious hazards to infrastructure. Furthermore, thaw of this ice can lead to extensive slope failures (Mackay, 1970; Lewkowicz, 2007), some of which can extend to many hectares in area (Lacelle et al., 2010). In addition to these hazards, the growth and thaw of near-surface ground ice can greatly influence landscape morphology, ecology and hydrology. Specifically, interactions between near-surface ground ice, vegetation and snow can substantially change the tundra landscape. The following sections describe the impacts of ground ice on subsidence, terrain stability and morphology, and interactions between different environmental components.

2.6.1 Ground subsidence degradation effect

When ice-rich permafrost melts, the ground surface subsides depending on the ice content (Mackay, 1970; Haeberli and Burn, 2002). Subsidence is already affecting engineered structures in the Arctic, and much of the infrastructure in northern regions is at high risk of subsidence under climate-change scenarios (Nelson et al., 2001). The damage that degrading permafrost may cause to infrastructure is demonstrated in Figure 2.8. Irregular terrain resulting from melted ice-rich ground and thaw settlement is collectively termed 'thermokarst' (Williams and Smith, 1989). Thermokarst development

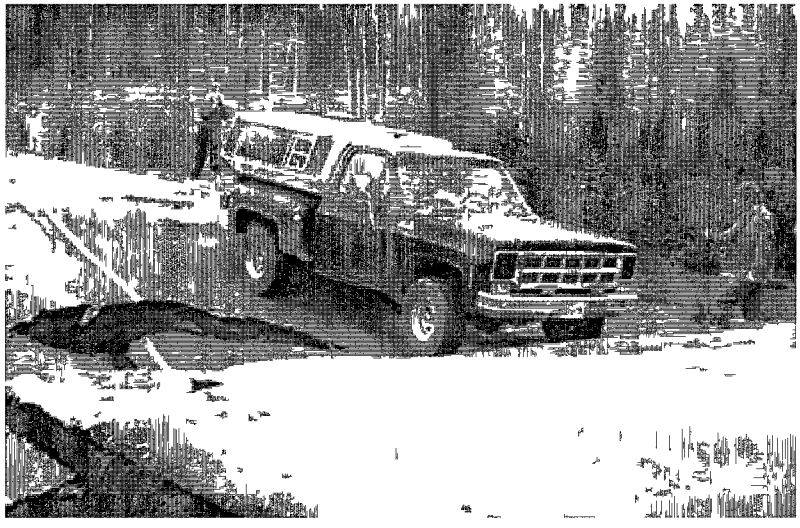


Figure 2.8 Ground subsidence caused by degrading permafrost, Alaska Highway (Turchetta, 2011).

follows climate warming or surface disturbances such as wildfires or vegetation removal, which both lead to changes in the surface energy balance and ultimately, thickening of the active layer (Burn and Smith, 1990).

Fire tends to cause rapid increases in active-layer depths over the first few years, from destruction of the vegetation cover leading to lower heat loss from evapotranspiration, a reduction in surface albedo, and removal of insulating mosses and peat (Mackay, 1995). In areas with ice-rich permafrost, the disturbance to the active layer leads to near-surface ground ice thaw and thermokarst subsidence (Mackay, 1995). Thaw of the active layer over summer can be described by the Stefan equation:

$$(2) \qquad z = \sqrt{\frac{2TK_T}{L}}$$

where z is the thickness of the active layer, T is the ground surface temperature, K_T is the thermal conductivity of unfrozen soil, and L is the volumetric latent heat of fusion. The increase in active-layer depths following fire results from greater surface temperatures in areas with disturbed ground. Downward thaw from the surface occurs rapidly at first because the thawed layer is initially thin and heat must travel through a thinner layer of unfrozen soil to warm the frozen ground below. As the thawed layer thickens, the progression of the thaw front slows because heat must travel through a greater thickness of unfrozen soil. Therefore, progression of the thaw front through the active layer decreases over time. As vegetation cover re-establishes following the disturbance, the albedo, evapotranspiration and insulating surface organic matter increase, the surface temperature decreases, and the active layer subsequently thins. Over time, the active layer

may thin to pre-fire depths as vegetation re-establishes. Ground uplift may occur from aggradational ice growth if the permafrost table rises (Mackay, 1995).

Conversely, if thermokarst depressions fill with water following disturbance to the ground surface, they may become small lakes. These lakes grow when further ground ice thaws, and the perimeter of the lake often slumps progressively as the disturbance extends outwards (Williams and Smith, 1989). The axial growth rate of thermokarst lakes in central Yukon has been reported up to 1.2 m/yr (Burn and Smith, 1990). Thermokarst lake development may cease following catastrophic drainage (Mackay, 1988), or by sediment infilling, and permafrost may eventually return to its pre-disturbance condition. Since the development of thermokarst usually involves ponds and lakes, it occurs in areas of low relief, because such bodies of water cannot exist on sloping ground (Williams and Smith, 1989).

Thaw settlement may also threaten low-lying, coastal areas by increasing the frequency of flooding as the ground surface settles. For example, subsidence of alluvial wetlands with substantial excess-ice content could contribute to more frequent and severe flooding events in the Mackenzie Delta area (Morse et al., 2009).

2.6.2 Slope stability

The stability of slopes in permafrost terrain is largely controlled by near-surface ground ice. Rapid thaw of ice-rich soils may lead to high pore-water pressures at the interface between thawed and frozen ground (Lewkowicz, 2007). This disturbance to the substrate significantly lowers the basal shear strength and may trigger slides, flows and slumps (Nelson et al., 2001; Lewkowicz, 2007).

Active-layer detachment slides are common slope failures in permafrost terrain resulting from the thaw of ice-rich soils (Lewkowicz and Harris, 2005; Lewkowicz, 2007). These features have a shallow failure plane parallel to the ground surface, and occur on gentle to moderate slopes (Lewkowicz and Harris, 2005). Sliding may occur almost instantaneously in some cases, while others enlarge over several days and flow at rates of a few metres per hour (Lewkowicz, 2007).

Although some active-layer detachment slides stabilize, further melting of near-surface ground ice in the scar floor may lead to the initiation of a retrogressive thaw slump (Lewkowicz and Harris, 2005). Retrogressive thaw slumps have a steep, exposed headwall of ice-rich sediment, capped by ice-poor overburden (Burn and Lewkowicz, 1990). Below the headwall, a low-angle mudflow area extends downslope. Once initiated, the slump progressively retreats due to further thawing of ground ice. Rates of headwall retreat may be several meters in a summer (Lantz and Kokelj, 2008). Thaw slumps stabilize when the supply of ground ice at the headwall is depleted, or when materials cover and insulate the ice from further melt (Burn and Lewkowicz, 1990).

Slope failures such as detachment slides and retrogressive-thaw slumps are hazardous to proximal infrastructure such as pipelines or roads. The rate of thaw slumping has increased recently in the Mackenzie Delta area, likely due to warmer and wetter summer conditions (Lantz and Kokelj, 2008; Lacelle et al., 2010). Continued interest in northern development and the changing climate conditions will likely present ongoing challenges to northern infrastructure from slope failures in ice-rich terrain.

2.6.3 Hummocks associated with aggradational ice growth

Hummocks are common in permafrost terrain (Mackay, 1980). These mound-shaped features, composed of fine-grained, frost-susceptible soils, may be vegetated (earth hummocks) or completely bare (mud hummocks) (Mackay, 1980). The permafrost table below hummocks is bowl-shaped and mirrors the hummock relief (Kokelj et al., 2007). Insulation from the growth of vegetation and accumulation of organic matter in hummock troughs results in shallow underlying active layers, whereas thicker active layers exist under less-vegetated hummock tops with more mineral soils. Ice lenses form parallel to the permafrost table, and are responsible for hummock growth as they thrust soils inward and upward (Kokelj et al., 2007). In contrast, the thaw of ground ice in hummocky terrain results in a decrease in hummock relief and diameter (Kokelj et al., 2007).

Changes in microrelief from thawing near-surface ground ice alter the hydrological regime and change the vegetation composition. For example, following thaw of ground ice at an Alaskan research site in tussock tundra, Osterkamp et al. (2009) observed wetter, more nutrient rich soil in newly developed thermokarst depressions, which enhanced tussock growth. Adjacent to thermokarst pits and gullies, the ground dried and tussocks were replaced with deciduous shrubs (Osterkamp et al., 2009).

Near-surface ground ice in hummocky terrain also influences vegetation structure in forested areas. Frost heave from aggradational ice growth has resulted in dramatically tilted spruce trees ('drunken forest') at sites in the Mackenzie River delta (Kokelj and Burn, 2003a). Sites with the most drastically tilted trees were underlain by ice-rich

permafrost, while study areas with straighter trees were underlain by permafrost with low excess-ice content (Kokelj and Burn, 2003a, Figs. 3 and 4).

2.6.4 Snow-shrub interactions

The above sections have dealt with geomorphic implications of near-surface ground ice. In addition to these effects, thaw of near-surface ground ice influences hydrology and vegetation growth. This section discusses relations between ground ice thaw, nutrient and water supply to the active layer, vegetation growth, and snow entrapment, and describes a possible feedback loop between these variables.

When near-surface ground ice melts following deep active-layer thaw, nutrients and water may be released to the active layer (Mackay, 1995). Rapid growth of vegetation, likely aided by increased nutrient and moisture supply, has been observed following a wildfire at sites near Inuvik, NWT (Mackay, 1995). The growth of tall vegetation in turn results in snow entrapment, as trees and shrubs reduce wind speed close to the ground and shelter downwind areas (Sturm et al., 2001). In addition, trapping of snow by vegetation reduces wind-driven sublimation, allowing the accumulation of the snow pack within and around vegetation.

Snow entrapment causes increased permafrost temperatures by acting as a buffer between cold air and the ground surface and limiting heat loss from the ground in winter (Sturm et al., 1997; Sturm et al., 2001; Mackay and Burn, 2002a). This winter warming dominates increased summer cooling from greater shading and insulation, and results in increased mean annual ground temperatures, which have been observed at Illisarvik (Mackay and Burn, 2002a). The increase in permafrost temperature may lead to

additional thaw of near-surface ground ice. The melted ice may further supply the active layer with nutrients and water (Mackay, 1995; Sugimoto et al., 2002). A positive feedback may result as water and nutrients promote further shrub growth, snow entrapment, and permafrost degradation (Sturm et al., 2001).

3. STUDY AREA AND METHODOLOGY

3.1 Introduction

This thesis examines the development of the ice-rich zone at the top of permafrost and relations between ice accumulation and controlling factors such as soil moisture, soil texture, temperature, and organic matter content. In addition, active-layer development at the study site is examined. Field data collection took place at the Illisarvik drained lake basin and surrounding tundra on northwest Richards Island, NWT. Samples of the upper metre of permafrost were obtained by drilling at 26 sites in the lake basin, nine others in the adjacent tundra and three on former lakeshore terraces. At each drill site, data on vegetation height, relative soil moisture status, and site topography were also collected. The active-layer depth at each site was estimated by probing in mid-August 2010. This chapter describes the location, history, and physical characteristics of the study site, and the sampling of near-surface permafrost. Laboratory and analytical methods are also presented.

3.2 Study area

3.2.1 Regional setting

The study area (Figure 3.1) lies within the Southern Arctic ecozone in the Tuktoyaktuk Coastlands physiographic region near the western Arctic coast (Rampton, 1988). Illisarvik is within the Kittigazuit Low Hills physiographic subdivision (Mackay, 1963). The surficial materials in Kittigazuit Low Hills are a range of sediments from fine-

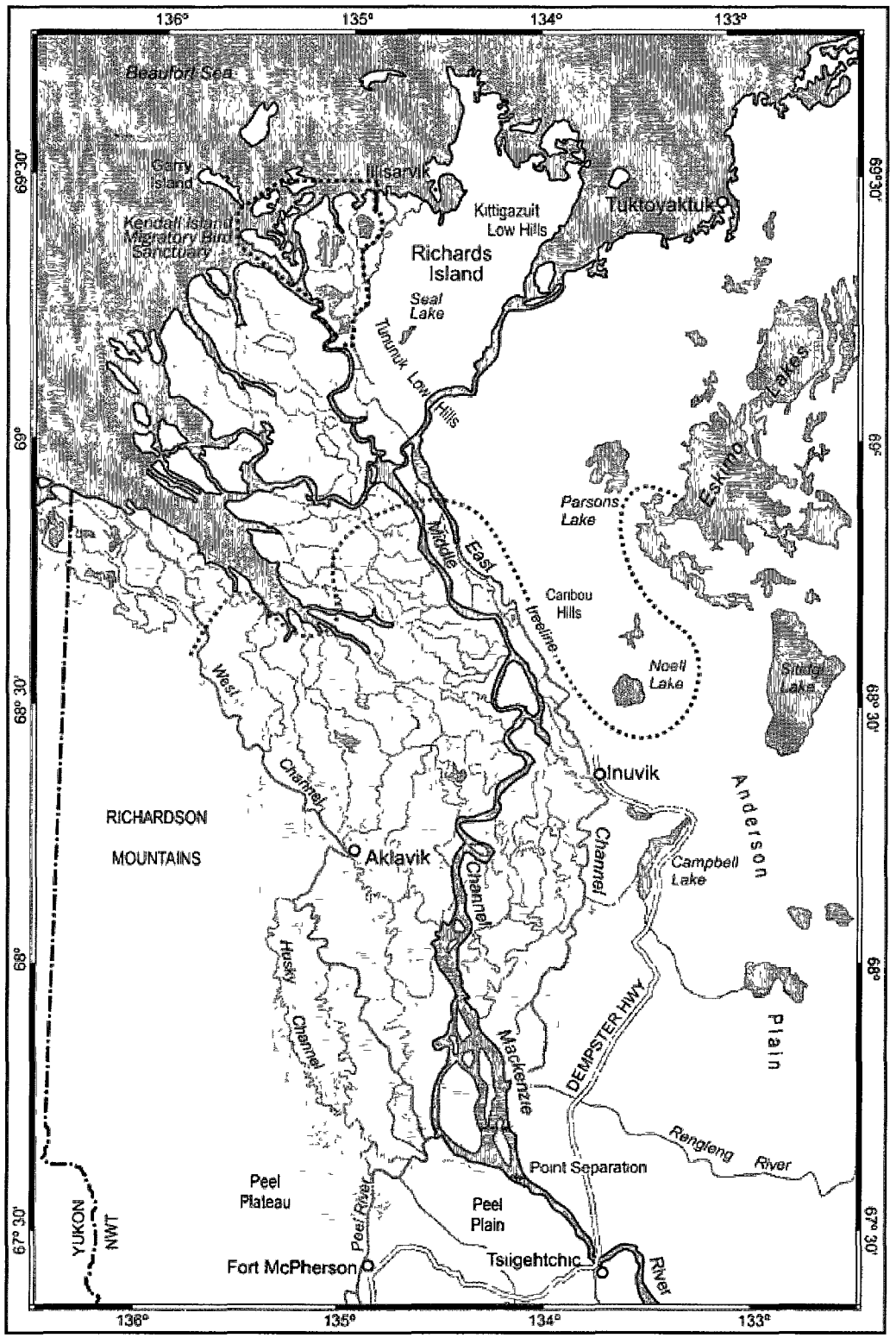


Figure 3.1 Location of study area (from Burn and Kokelj, 2009). Illisarvik is marked in the top centre of the image.

grained till to glaciofluvial gravel (Burn, 2002). The topography of the western side of Richards Island is irregular, with rolling hills and many poorly-drained flats and drowned valleys (Mackay, 1963). Earth hummocks commonly cover the tundra terrain (Mackay, 1963)

The shrub tundra vegetation in the area is characterised by sedges and dwarf shrubs such as willow (*Salix* spp.), ground birch (*Betula nana*) and green alder (*Alnus crispa*) (Mackay, 1963). Poorly drained areas are dominated by sedges, while drier upland sites host grasses and shrubs (Mackay, 1963). Grasses and lichens grow on hummock tops, while mosses and shrubs are found in interhummock depressions. Tundra vegetation at the study site is shown in Figure 3.2.

3.2.2 Climate

Western Arctic Canada has long, cold winters and short, cool summers. Tuktoyaktuk is the nearest community to the study area, located about 60 km east on the coast of the Beaufort Sea. The annual mean air temperature at Tuktoyaktuk is -10.6°C (Environment Canada, 2011). July is usually the warmest month (11°C), and January is typically coldest (-27°C) (Figure 3.3). Annual precipitation at Tuktoyaktuk is relatively low (170 mm), and the majority falls in the summer and autumn (Figure 3.3).

3.2.3 Permafrost

The study site is situated in the continuous permafrost zone (Heginbottom et al., 1995). Richards Island was last glaciated sometime between ~22,000 and 16,000 years BP (Rampton, 1988; Murton, 2009). The recent absence of glacial ice has resulted in

Figure 3.2 Upland tundra vegetation around the Illisarvik drained lake basin.

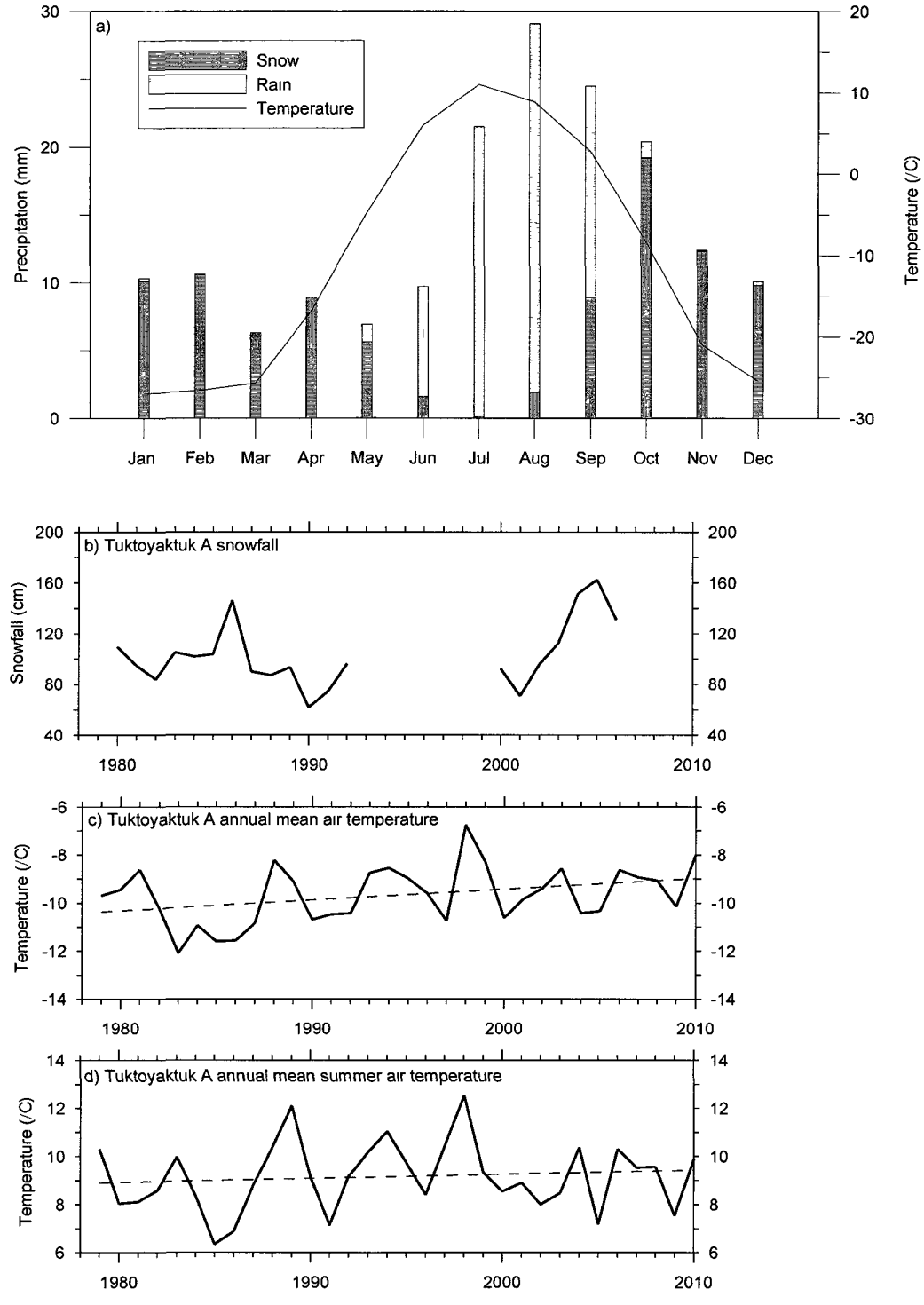


Figure 3.3 (a) Monthly climate normals (1971-2000) for Tuktoyaktuk A, (b) available snowfall record for Tuktoyaktuk A, c) annual mean air temperature for Tuktoyaktuk A, and d) annual mean summer air temperature for Tuktoyaktuk A (Environment Canada, 2011). Dashed lines in the temperature records represent the line of best fit by least squares linear regression.

permafrost thicknesses generally over 400 m (Judge et al., 1987). Permafrost temperatures in the region range from -6 to -7°C (Burn and Kokelj, 2009).

The near-surface permafrost on Richards Island is commonly ice-rich, mainly due to wedge, segregated, and pore ice (Pollard and French, 1980). In the top 10 m of permafrost, it is estimated that 14.3% of the volume is occupied by excess ice, mainly due to ice wedges (Pollard and French, 1980). The large amount of ice-rich permafrost results in an abundance of thermokarst activity. Thermokarst is most evident by the abundance of tundra lakes on Richards Island. The over 1200 thermokarst lakes on Richards Island cover 23.5% of the ground surface (Burn, 2002). Taliks, or unfrozen zones, exist beneath lakes that do not freeze through in the winter. Mackay (1988) determined that one or two tundra lakes along the western Arctic coast drain naturally every year.

3.3 Lake Illisarvik

The research site is the Illisarvik experimentally drained lake basin (see Mackay, 1997). Illisarvik is located at 69°28'50.76" N, 134°35'04.32" W. Prior to drainage, Lake Illisarvik measured approximately 600 x 300 m, and was underlain by a bowl-shaped talik 32 m deep in the lake centre (Mackay and Burn, 2002a). The lake formed approximately 9500 years ago, and quickly expanded until reaching its maximum size around 6000 years ago (Michel et al., 1989). Its contemporary size was reached near 2000 years ago after it partially drained (Michel et al., 1989). Lake depths generally ranged between 2 and 3 m before drainage (Mackay and Burn, 2002a). Dr. J.R. Mackay

drained the lake on 13 August 1978 in order to initiate and study permafrost growth (Mackay, 1997). Two residual ponds were left in the lake bed following drainage, termed North Pond and South Pond (Mackay and Burn, 2002a). South Pond, near the centre of the lake bed, is 4 m deep.

Illisarvik is Canada's longest-running northern field experiment. Studies have been ongoing at the site for the past three decades (e.g. Burgess et al., 1982; Michel et al., 1989; Mackay, 1997; Mackay and Burn, 2002a; Burn and Zhang, 2010). Prior to drainage, ground temperatures, the position of the permafrost table, and the composition and DC resistivity of basin sediments were main topics of research (e.g. Hunter et al., 1980; Scott, 1980). Following drainage, research continued on ground temperatures, permafrost growth, active-layer development, ground water isotopes, frost heave, ice-wedge growth, vegetation, and aggradational ground-ice development (see Burn and Burgess, 2000).

3.3.1 Permafrost aggradation

Permafrost began aggrading into the basin talik during the first winter following drainage (Mackay and Burn, 2002a). Twenty-four temperature cables were installed in boreholes in the lake bottom in order to monitor ground temperatures beneath the lake and surrounding shorelines (Burgess et al., 1982). Two years after drainage, the talik had completely frozen back where it had been less than ten metres thick at nearshore sites (Burgess et al., 1982). In the lake centre, five or six metres of permafrost had formed after the first two years (Burgess et al., 1982). Mackay (1997) presented a cross-section showing the progression of the 0°C isotherm from the northeast margin of the lake

towards the lake centre between 1978 and 1983 (Figure 3.4). By five years following drainage, the 0°C isotherms were nearly converging in the lake centre, where the pre-drainage talik was deepest. At Benchmark 63 in the middle of the lake basin, the ground was below 0°C to the previous depth of the talik by the mid-1980s (Mackay 1997, Fig. 14), indicating that permafrost completely established in the basin sediments within a decade following drainage. However, there is no permafrost immediately beneath South Pond.

3.3.2 Lake basin soils

The lake basin sediments can be divided into three distinct units, which are overlain by peat in some areas (Michel et al., 1989). The deepest unit is a thick layer of fine-medium sand that contains some organic fragments and is likely deltaic in origin (Michel et al., 1989). Interbedded layers of sand and clay-silt exist above this unit. A middle unit of clay-silt of variable sand content lies above the interbedded layers (Michel et al., 1989). This unit is low in organics and contains many small stones. The uppermost unit consists of organic-rich silts. The unit is thickest in the centre of the lake basin, and thins towards the margins where sands dominate (Michel et al., 1989).

3.3.3 Ground ice growth

Since drainage, segregated ice and wedge ice has formed in the basin. Mackay and Burn (2002a) monitored aggradational ground ice growth up to 1999 by measuring the ground surface rise beneath benchmark brackets throughout the lake basin (Figure 3.5). The ground level at several benchmarks in the north end of the lake basin, where willows are abundant, rose between 1980 and 1987, indicating ice accumulation at the

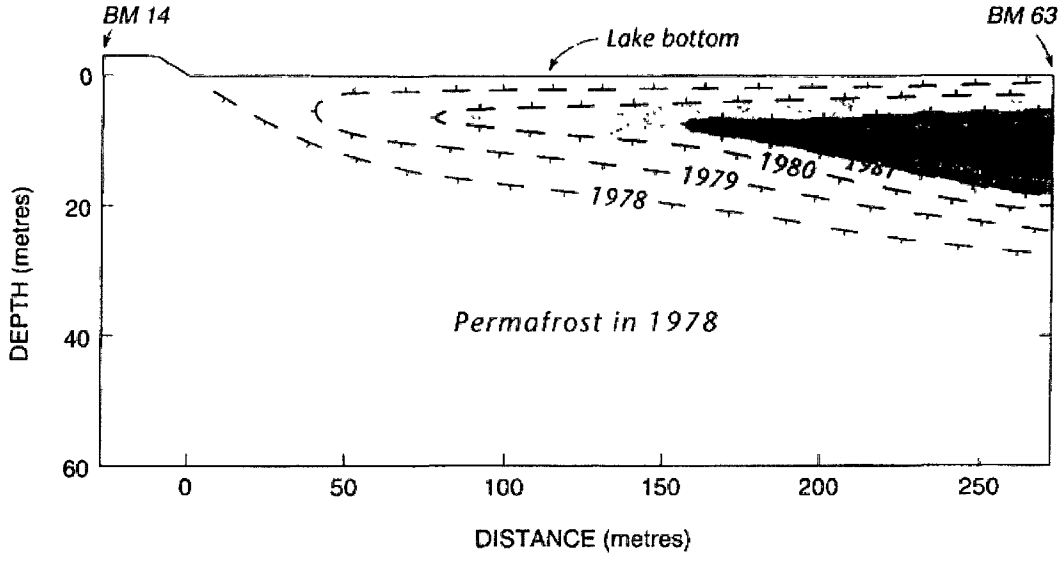


Figure 3.4 The progression of 0°C isotherms from the northeast lake edge towards the lake centre between 1978 and 1983 (Mackay, 1997, Fig 11). The black shading represents unfrozen ground. BM 14 is a benchmark on the lake shore and BM 63 is located in the lake centre.

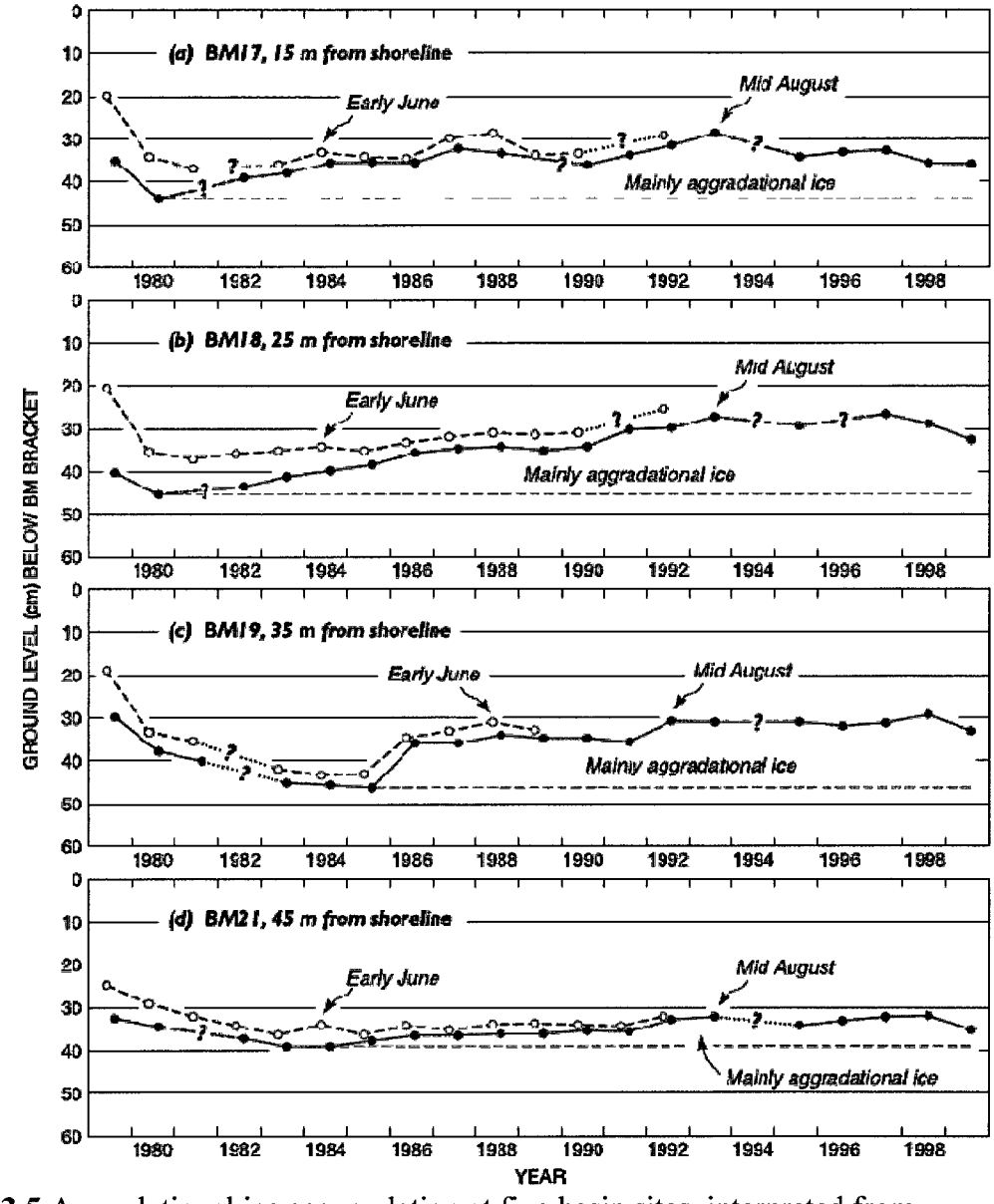


Figure 3.5 Aggradational ice accumulation at five basin sites, interpreted from benchmark protrustions (Mackay and Burn, 2002a, Fig. 15).

top of permafrost (Mackay and Burn, 2002a). Subsidence occurred between 1987 and 1990, likely due to summer thaw of ice. From 1990 to 1993, the ground surface rose once again, and between 1993 and 1999, the surface subsided coincidently with a thickening active layer (Mackay and Burn, 2002a). In total, between 1980 and 1999, up to 10 cm of ice grew at the top of permafrost. This amount corresponded with results obtained by drilling at six sites in the basin (Mackay and Burn, 2002a, Fig. 17). The growth of ground ice at these sites is not likely representative of the entire lake basin as they are situated in very moist ground with frost-susceptible soils, conditions that may be more favourable to ice formation than in other areas in the basin.

Ice-wedge growth began in the lake basin during the first winter following drainage (Mackay, 1986). Thermal contraction crack widths observed at the surface ranged between 3 and 20 cm, and ice-wedge growth during the first 10 years was between about 1 and 3 cm yr^{-1} (Mackay and Burn, 2002b). However, increases in ground temperatures from vegetation growth and snow entrapment caused the cessation of thermal contraction cracking by 12 years after drainage (Mackay and Burn, 2002b).

3.3.4 Active-layer development

Active-layer depths at Illisarvik are controlled by a number of factors including permafrost temperatures, vegetation, soils, soil moisture, and exposure (Mackay, 1982). At the start of the first summer following drainage of Lake Illisarvik, active-layer transects were established along the two lake axes (Figure 3.6). Between 1979 and 1981, as the permafrost aggraded, variations in active-layer thickness were similar in magnitude to sites with old permafrost (Mackay, 1982). Windblown accumulation of organic matter

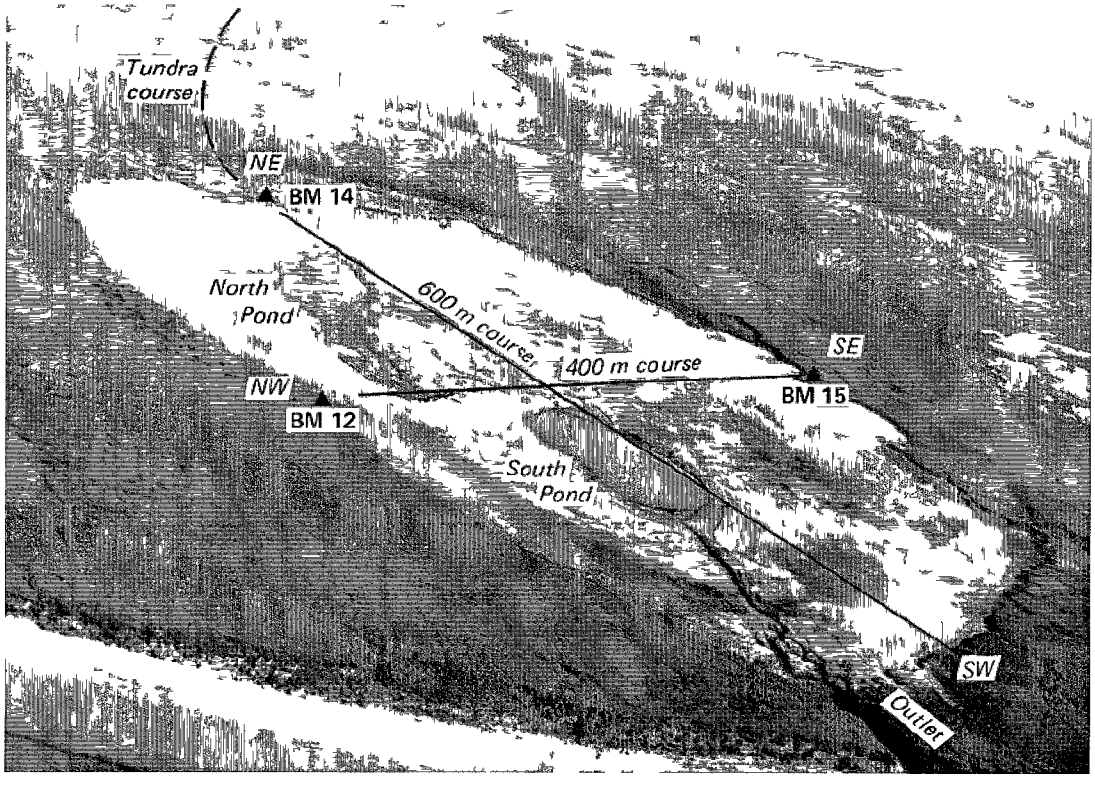


Figure 3.6 Illisarvik on August 12, 1979, one year following drainage. The active-layer transects along the major and minor axes of the basin are marked (modified from Mackay and Burn, 2002a, Fig. 2).

at some sites during this period caused the active layer to vary greatly from year to year (Mackay, 1982). Vegetation had not established at any active-layer transect sites by 1981, and so it was not a factor in the early development of active-layer depths. The lack of vegetation also meant that winter snow accumulation was minimal (Mackay, 1982).

Mackay and Burn (2002a) monitored active-layer depths along the lake axes until 1999. Along the lake margins, the active layer was >90 cm between 1979 and 1999. This is attributed to the well-drained, sandy soils, and insulation from deep snow banks in the winter. In other areas of the basin, active-layer depths generally ranged between 50 and 60 cm from 1979-1988 (Mackay and Burn, 2002a). Between 1989 and 1993, the active-layer depths gradually increased, and by 1994-1999, the basin depths were on average 20% greater than during 1979-1983. This increase was due to the establishment of vegetation and resulting snow entrapment leading to increased ground temperatures (Mackay and Burn, 2002a). In the lake basin, site-specific effects like soil type and snow accumulation controlled the active-layer depths. Data from twelve sites along a 550 m long active-layer transect in the tundra adjacent to the drained lake basin indicated that summer temperature is the controlling factor at these sites (Mackay and Burn, 2002a).

3.3.5 Lake basin vegetation succession

Initially, the newly-drained lake bottom was covered in places by a layer of wave-eroded surface peat (Mackay and Burn, 2002a). By summer 1981, some of the peat surface was eroded by strong winds and deposited in a thick layer at the southern margins of the lake bottom (Mackay, 1982). Mackay and Burn (2002b) described the vegetation at four ice-wedge monitoring sites (Figure 3.7). In the summer of 1980, vegetation spread significantly, mainly in the organic silts around North Pond (Figure 3.7, 'Site 1')

Figure 3.7 Illisarvik on August 12, 1979, one year following drainage. Four ice-wedge monitoring sites are identified (modified from Mackay and Burn, 2002b).

(Mackay, 1986; Mackay and Burn, 2002b). Marsh fleabane (*Senecio congestus*) grew rapidly adjacent to North Pond, and by the 1981-1982 winter, snow was being trapped by the vegetation to depths of about 60 cm (Mackay, 1986). By the summer of 1983, grasses, sedges (*Cyperaceae*), and horsetails (*Equisetum* spp.) covered about two thirds of 'Site 1' (Mackay and Burn, 2002b). At 'Site 3', south of North Pond, vegetation establishment was slower in the dryer, sandier soils. In 1983, scattered clumps of grasses were growing a few metres apart at 'Site 3' (Mackay and Burn, 2002b).

Ovenden (1986) conducted a detailed vegetation survey of the lake bottom in summer 1985. The lake bottom vegetation varied in composition and distribution, and was controlled predominantly by surface moisture conditions and proximity to the former shoreline (Ovenden, 1986). The marginal areas likely receive more moisture than interior areas of the lake basin, from downslope seepage and snow beds (Ovenden, 1986; Mackay and Burn, 2002a).

Seven years after drainage, vegetation in the basin was dominated by alkali grass (*Puccinellia borealis*), water sedge (*Carex aquatilis*), marsh fleabane, mayweed (*Matricaria ambigua*), and northern tansy mustard (*Descurainea sophioides*) (Ovenden, 1986). Four distinct vegetation units were identified (Figure 3.8): (1) areas within 10 m of the shoreline, (2) non-eroded areas further than 10 m from the shoreline, (3) eroded lake bed, and (4) wet depressions (Ovenden, 1986). The nearshore margin had the most continuous vegetation cover in 1985, and several species including willows were in higher abundance near the shoreline. Non-eroded areas of the lake bottom were 40-90% covered by up to 20 different plant species, and alkali grass dominated (Ovenden, 1986). Areas that experienced significant wind erosion of material were sparsely vegetated, and

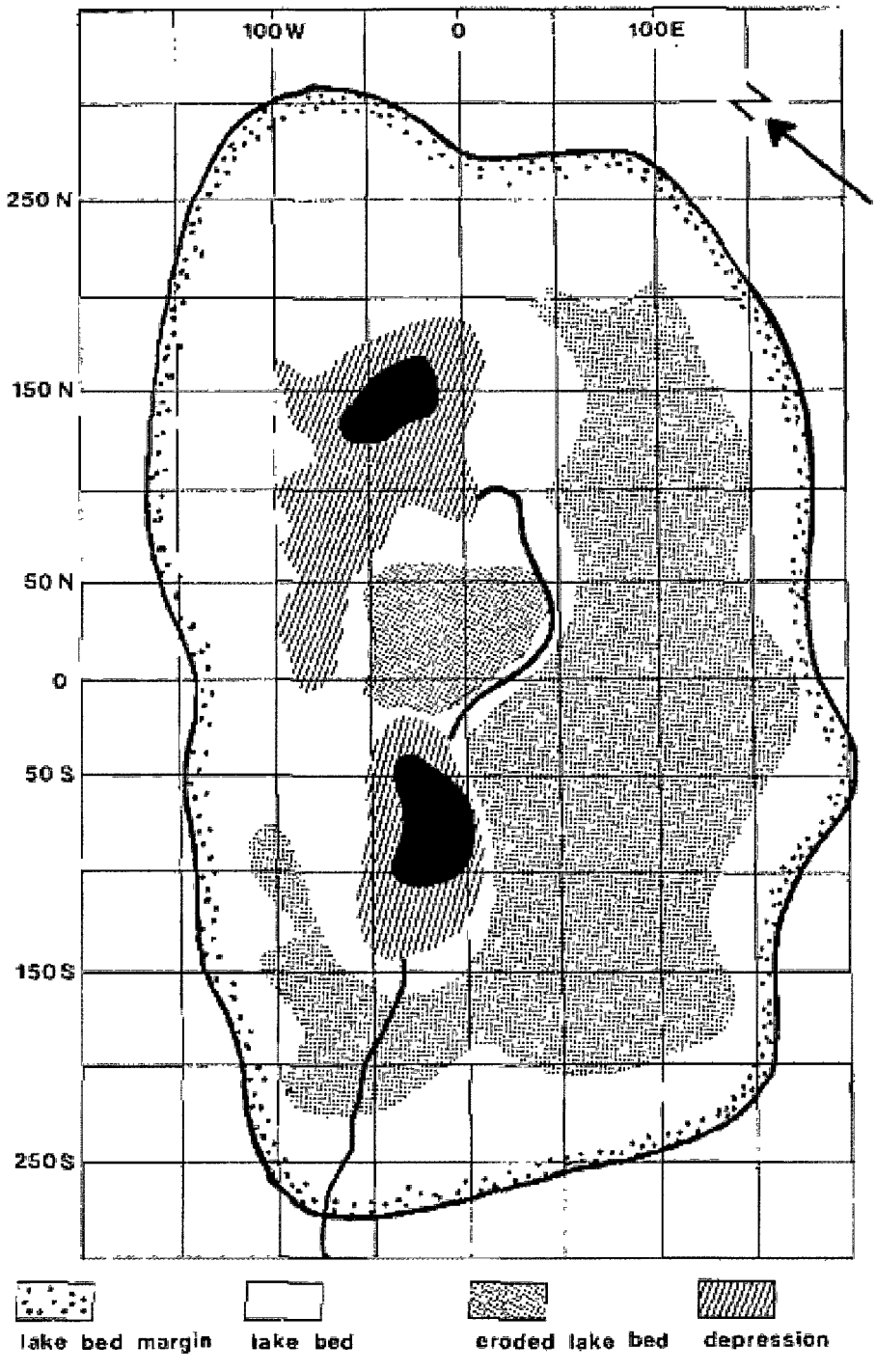


Figure 3.8 Vegetation classes in the lake basin established by Ovenden (1986, Fig. 2). Black areas represent water bodies.

wet depressions were bordered by marsh fleabane and other moisture-tolerant species (Ovenden, 1986).

Between 1985 and 1999, vegetation cover greatly increased in the basin (Figure 3.9). Grasses and low cover proliferated, and by 1999, some taller willows had established, mainly in the north and northeast of the basin and along lake margins (Mackay and Burn, 2002a). Between 1990 and 1999, the cover at 'Site 3' increased from 80% to 100%. Vegetation growth was slower at 'Site 2' than at 'Site 1' (Figure 3.7), but by 1999 vegetation was established, alders were growing, and some willows exceed 2 m in height (Mackay and Burn, 2002b). The area adjacent to south pond remained bare of vegetation due to frequent flooding during snowmelt and possible high salt concentrations from pore-water expulsion (Figure 3.10) (Mackay and Burn, 2002a).

Vegetation has continued to grow in the lake basin. In summer 2010, tall willows were abundant around North Pond and the lake margins, many in excess of 3 m in height. In the basin centre, grasses dominated the vegetation cover, and patches of willows were scattered. Vegetation remained sparse near the southern margins of the lake and adjacent to South Pond.

3.4 Tundra

Much of the rolling tundra at Illisarvik is covered by earth hummocks. These features are fully vegetated and composed of fine-grained, frost-susceptible soils (Mackay, 1980). Earth hummocks are circular or oval-shaped, up to 60 cm in height, and 100 to 200 cm in diameter (Tarnocai and Zoltai, 1978). The permafrost beneath

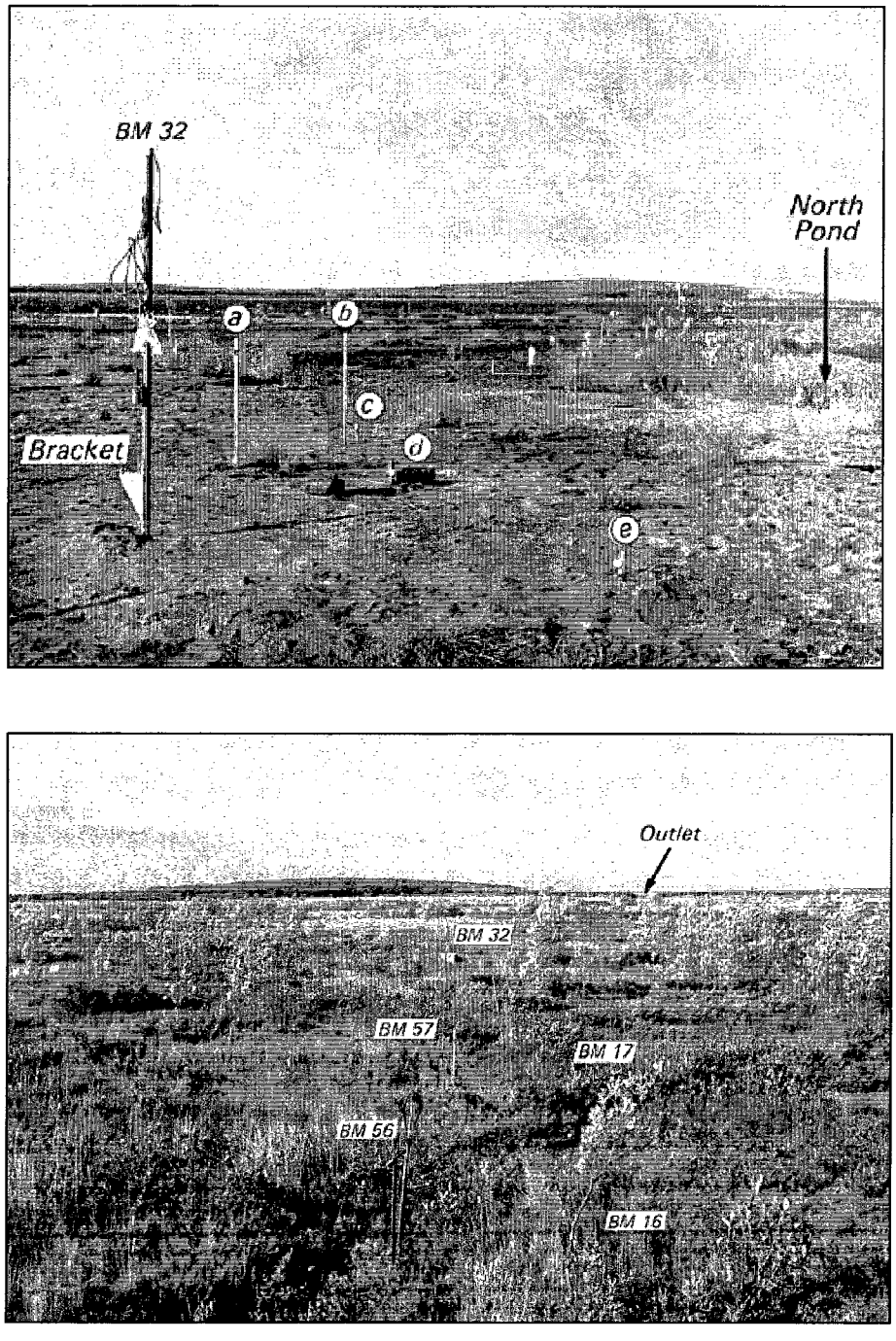


Figure 3.9 Drained lake bottom in summer 1981 (top) showing benchmark 32 surrounded by sparse vegetation, and in summer 1999 (bottom) showing abundant vegetation and willow growth (Mackay and Burn, 2002a, Fig. 4 and 5).

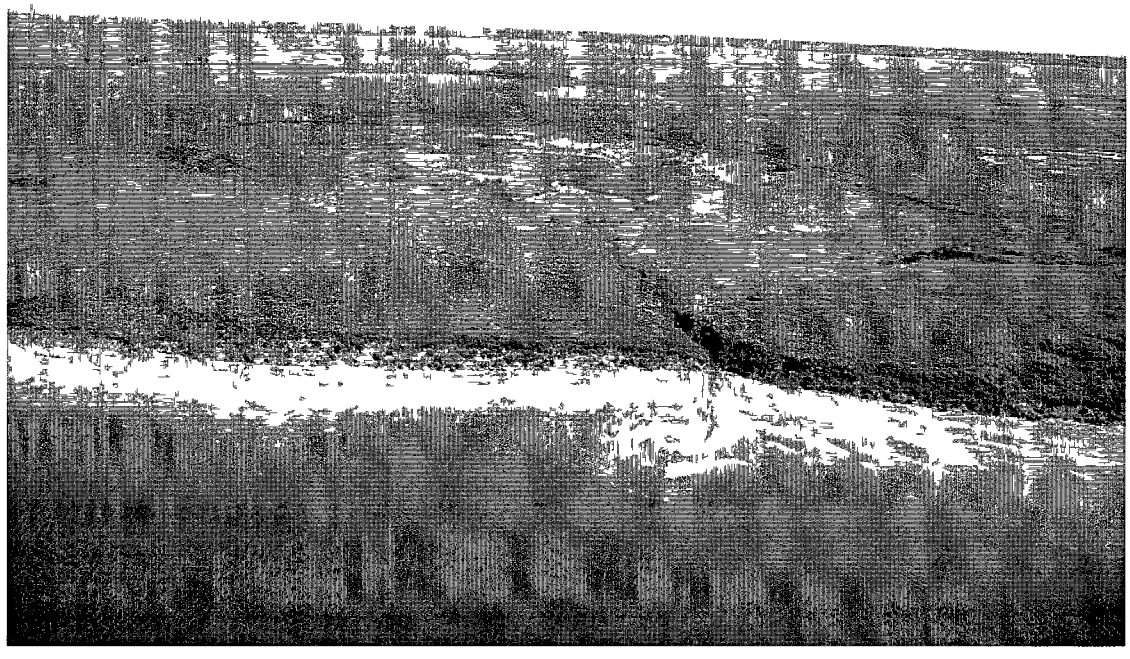


Figure 3.10 The drained lake basin in 2008. Note the increase in vegetation cover compared to Figure 3.7. Photo: C.R. Burn.

hummocks is characteristically ice-rich (Tarnocai and Zoltai, 1978; Mackay, 1980; Kokelj and Burn, 2003b).

The hummocks in the Illisarvik tundra are hundreds to thousands of years old (Mackay and Burn, 2002a). Mackay and Burn (2002a) installed stakes in hummock tops along a transect in the Illisarvik tundra. Between 1983 and 1999, the stakes tilted very little, indicating that the tundra hummocks are relatively inactive.

The topography of the tundra surrounding the lake basin leads to a variety of slope, aspect, moisture, and vegetation conditions. Higher, flat areas appear better drained compared to some sloped and low-lying locations. Vegetation in the tundra is described in section 3.2.1, and can be seen in Figure 3.2.

3.5 Study design

The Illisarvik drained lake basin is an ideal site to study the ice enrichment of upper permafrost for several reasons. First, permafrost began aggrading into the sub-lake talik following drainage, so ground ice growth has occurred in a temporally controlled setting. The ground ice conditions in the basin can be compared to those in the surrounding tundra, where accumulation in near-surface permafrost has likely occurred over the past 8000 years since the early-Holocene climatic optimum (e.g. Burn, 1997). Therefore, the control of long-term active layer history on ice enrichment can be examined. Second, the long research history of Illisarvik makes the site ideal for this investigation. The ground ice contents measured in this study can be compared to those obtained in past studies by Mackay and Burn (2002a) and Michel (1982). Long-term

measurements of active-layer depths and snow depths along the major lake axes are available for the analysis. These data provide spatial control to examine controlling factors of ice accumulation and active-layer variations. Finally, the data from this study will be of use to future ground ice investigations.

3.5.1 Site selection

Three visits were made to Illisarvik between June 24 and August 15, 2010 for sample collection. Twenty-six cores of near-surface permafrost were collected from the Illisarvik lake basin. In addition, nine cores were obtained from the surrounding tundra and three from former lakeshore terraces. Locations of sites within a previously-established grid system at Illisarvik are shown in Figure 3.11. The site selection reflected the goals of the study: to examine the relations between possible controlling factors on the development of near-surface ground ice, to compare present ice conditions to those recorded in past studies, and to compare ice conditions in the drained-lake basin to those in the tundra. Therefore, basin sites were chosen over a range of moisture, vegetation, and soil conditions, and adjacent (within 1.5 m) to previously-drilled locations. Tundra sites were selected adjacent to previously-drilled locations and near the tundra active-layer transect (see Mackay and Burn, 2002a).

3.5.1.1 Previous drill sites

Research by Michel (1982) and Mackay and Burn (2002a) included moisture content and ice accumulation data from the Illisarvik basin and surrounding tundra. Michel (1982) extracted cores in 1979 and 1980 from locations in the basin, tundra and former lakeshore terraces. Comparisons of gravimetric moisture content profiles were possible at two basin sites, one tundra site and one lake terrace site. Aggradational

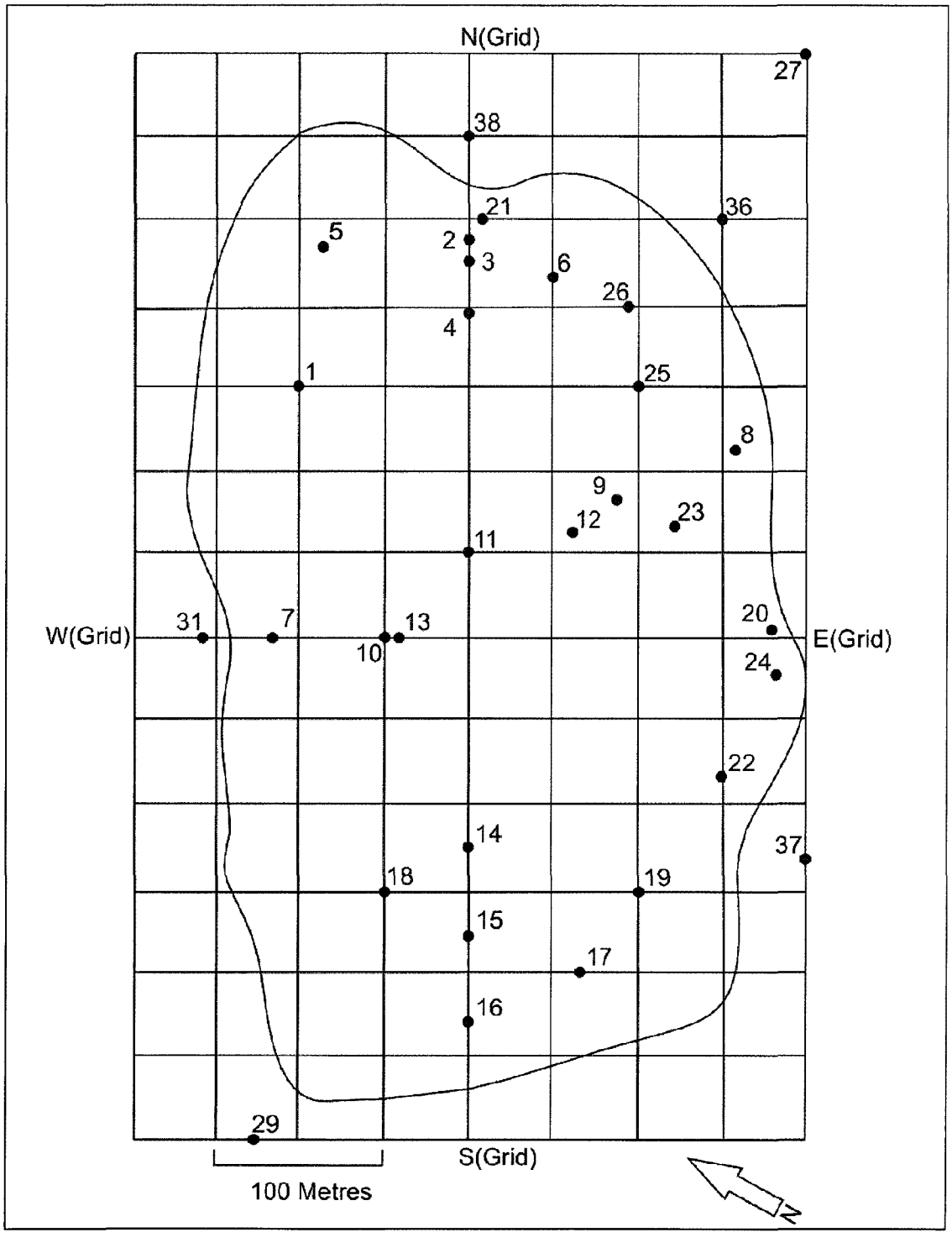


Figure 3.11 Locations of drill sites within a grid reference set up following drainage. The grid system is oriented along the major and minor lake axes, and stakes are set up 50 m apart . Six tundra sites drilled in this study outside the grid are not shown on this figure. The major and minor axes transects are labelled with the cardinal grid directions.

ice accumulation was determined by Mackay and Burn (2002a) at sites around North Pond, and results from this study are used as a comparison.

3.5.1.2 Basin transects

Eleven sites in the basin were located along transects established on the major and minor lake axes (Mackay and Burn, 2002a). This enabled analysis of the ground ice data in conjunction with active-layer and snow depths.

3.5.1.3 Soil moisture

Among the basin sites of this study, 13 were located in 'wet' areas and 13 in 'dry' areas. Drill sites were characterized as 'wet' when the base of the active layer was visibly saturated (where water drained from the active layer into the bore hole), there was standing water within 3 m, or by the growth of *Equisetum* spp. (horsetail) at the site. Horsetail grows in poorly drained, or moisture-rich areas (Aiken et al., 2001), and therefore may be used as an indicator of average soil-moisture status. These criteria were used as opposed to direct measurement at the time of drilling because the soil moisture of the active layer depends on the timing and intensity of precipitation events. Since the drilling occurred over three visits, active-layer moisture status at the time of drilling would not be comparable between different sampling times. Therefore, these indirect classification criteria were employed in an attempt to ascertain persistent moisture conditions. The selection of these sites enabled the influence of soil moisture on ice accumulation to be examined between the two distinguished classes.

3.5.1.4 Temperature

To assess temperature as a possible control on ice accumulation in the lake basin, sites were chosen over a range of vegetation conditions. Vegetation traps snow, which increases the temperature of permafrost (Sturm et al., 2001; Mackay and Burn, 2002a). Illisarvik basin snow data collected by C.R. Burn were used with vegetation heights measured in this study, and there was a significant statistical relation between vegetation height and snow depth ($r^2 = 0.66$; $p = 0.004$) (Figure 3.12). Warmer permafrost allows greater upward moisture migration away from the top of permafrost in winter, so less near-surface ground is expected in well vegetated areas. Therefore, sites with abundant vegetation growth and sites with little or no vegetation were of interest, in order to investigate whether permafrost temperature is related to near-surface ice enrichment in the basin.

3.5.1.5 Soil texture

Soils in the basin are variable both with depth and location in the basin (Michel et al., 1989). To capture a range of soil conditions in the sampling scheme, sites were established near the margins, where sands often dominate (Michel et al., 1989), and towards the centre of the basin where silts are more prevalent. This sampling strategy enabled the influence of soil texture on ice enrichment to be examined.

3.5.1.6 Limitations of site selection

There are several noted limitations in the sampling scheme. First, the classification between wet and dry sites was based on qualitative criteria. The moisture status of the active layer could be assessed in greater detail by instrumenting sites to obtain quantitative measures of the moisture status over the entire summer. As this was

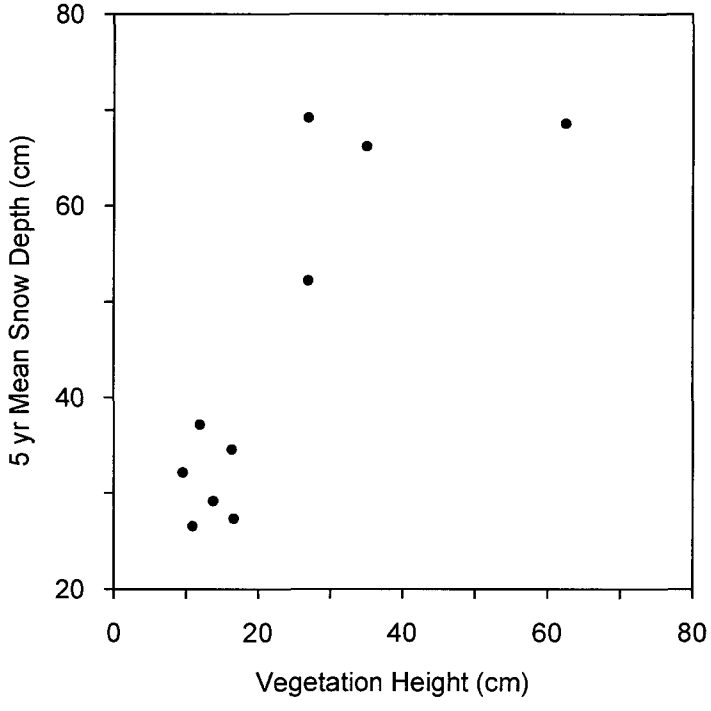


Figure 3.12 Vegetation height and mean snow depths (2005-2010) for 10 basin sites adjacent to benchmarks with snow depth records ($r^2 = 0.66$; $p = 0.004$).

beyond the scope and study period of the project, the qualitative assessment was considered an appropriate alternative.

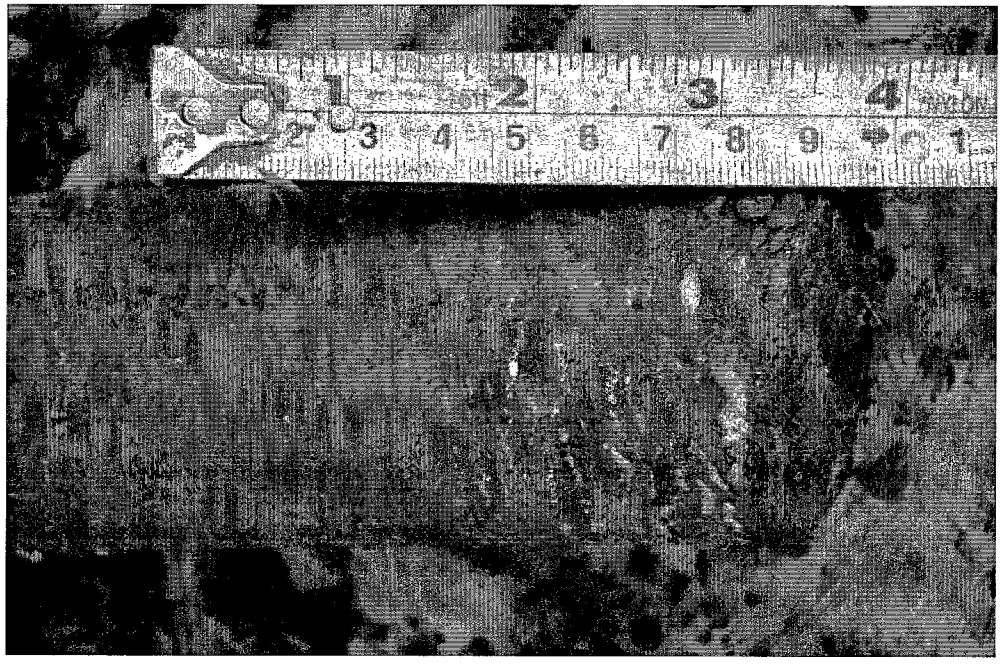
Problems arose while attempting to drill at many sites, both in densely willowed and bare areas. Some willow sites were particularly poorly drained, especially surrounding North Pond, and when attempting to drill, water and saturated sediment from the base of the active layer flowed into the borehole. Some drill holes had to be abandoned as a result. Many of the most bare areas at Illisarvik are around the eastern (grid) margins of the lake basin. The substrate in these areas is commonly sandy, and thaw depths are considerable because there is no insulating vegetation. This allows water to accumulate at the base of the thawed zone. When drilling was attempted at these sites, thawed, saturated sand would collapse into the bore hole. Due to these difficult conditions, the range of vegetation, moisture, and soil conditions sampled does not fully encompass those observed in the lake basin.

3.6 Field methods

3.6.1 Core extraction

Cores of near-surface permafrost were obtained using a CRREL drill (Figure 3.13). The core barrel retrieves samples 5 cm in diameter. The core segment length was determined using a measuring tape after it was removed from the barrel. The depth to the bottom of the borehole was measured after each segment was removed. This was accomplished by placing a dowel across the borehole, sliding a measuring tape to the bottom, and recording the length to the bottom of the dowel. The bottom of the dowel

a)



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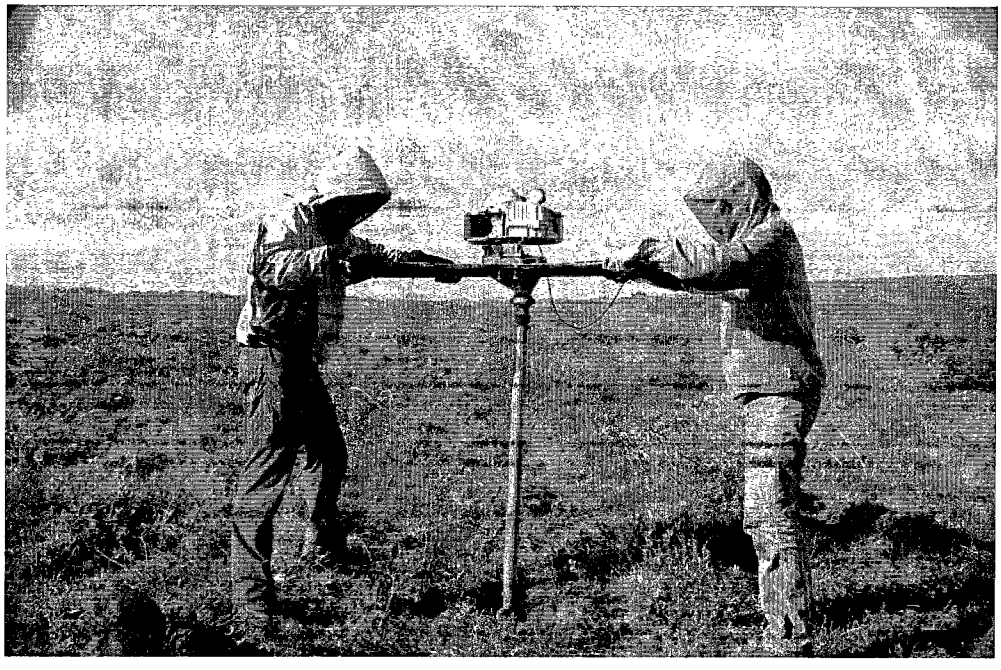


Figure 3.13 (a) Ice-rich permafrost core; (b) field use of CRREL drill.

was taken to be the ground surface. In some cases, when drilling to deeper depths, a graduated steel rod was used for this measurement.

Core segments of 10 cm length were cut using a hammer and chisel. General characteristics of each section such as soil type, colour, ice content, and visible organic matter were recorded. These sections were then placed in plastic sample bags. Later, the samples were double-bagged and weighed in the field so that they could be re-weighed in the Inuvik lab to detect any leakage. The objective was to sample the top metre of permafrost; however, many cores were longer than this, and due to obstructions and the occasional collapse of boreholes, some were shorter.

Each drill site was photographed for reference, and general site characteristics were recorded. These included the topography, vegetation and moisture conditions. If a distinct organic horizon existed at the surface, this was measured with a tape.

3.6.2 Vegetation height

Vegetation height in this thesis refers to the mean of 8 measurements of maximum vegetation height within a 2 m radius of the drill site. The maximum height refers to the upper extent of foliage that has the potential to trap snow. A 2 m radius was used for the measurements because vegetation can affect the snow depth and ground temperature of the surrounding area (Sturm at al., 2001).

3.6.3 Active layer

Thaw depth was probed at the time of drilling, and in mid-August 2010 to estimate the thickness of the active layer. The thaw depth was measured with a graduated steel probe pushed into the ground to the depth of refusal. Three measurements were

taken immediately adjacent to the borehole, and the maximum depth was used as an estimate of the active-layer depth.

3.6.4 Measurement uncertainties

There are a number of acknowledged uncertainties associated with field measurements in this study. First, sections of core were measured following extraction, and the length compared with the borehole depth measurements. There were often discrepancies between the measured core length and the depth to the bottom of the borehole. Core segments were commonly longer than the measured depth of the borehole, indicating that the core may expand following extraction to the surface. Possible reasons for this expansion are the removal of overburden pressure and/or thermal expansion as the core is brought to higher temperatures. Ice-rich core segments often broke up during extraction at locations with larger ice lenses, making accurate length measurements impossible. When discrepancies between measurements existed, or when the core segments were ice rich, the borehole depth was used to obtain the core segment length.

The accuracy of borehole depths using the measuring tape or graduated steel probe are thought to be within ± 1 centimetre. Error in these measurements may be due to the measuring tape bending in the borehole, or the probe not being perpendicular to the ground surface during measurement. Bending of the tape was minimized by lowering it until the bottom of the hole was firmly located, and then by limiting the pressure on the tape.

Mackay (1982) reported that estimates of active-layer depths at Illisarvik are accurate to several centimetres for shallow (e.g. 20 to 50 cm) depths, and perhaps to 10

cm for depths greater than 100 cm. The accuracy may be better for the current study because near-surface ground ice is present at the top of permafrost at many sites, and offers resistance while probing. The active-layer measurement error affects the accuracy of, for example, moisture and excess ice contents plotted against the depth below the permafrost table.

3.7 Laboratory and analytical methods

3.7.1 Gravimetric moisture content

Gravimetric moisture content was determined for each core segment. The samples were weighed in the lab to ensure no leakage occurred during storage in the field and transportation. The samples were placed in an oven and dried at 105°C. The dried samples were reweighed in order to calculate the gravimetric moisture content:

(3) $$GMC = W_m/S_m$$

where GMC is the gravimetric moisture content, W_m is the mass of water, and S_m is the mass of dried sediment.

3.7.2 Excess-ice content

The excess-ice content was determined for each ice-rich 10 cm sample segment. Ice-rich samples were homogenised, poured into beakers, weighed and allowed to settle (Figure 3.14) (e.g. Kokelj and Burn, 2003). The volumes of saturated sediment and supernatant water were recorded. The excess-ice content was estimated using (Kokelj and Burn, 2003):

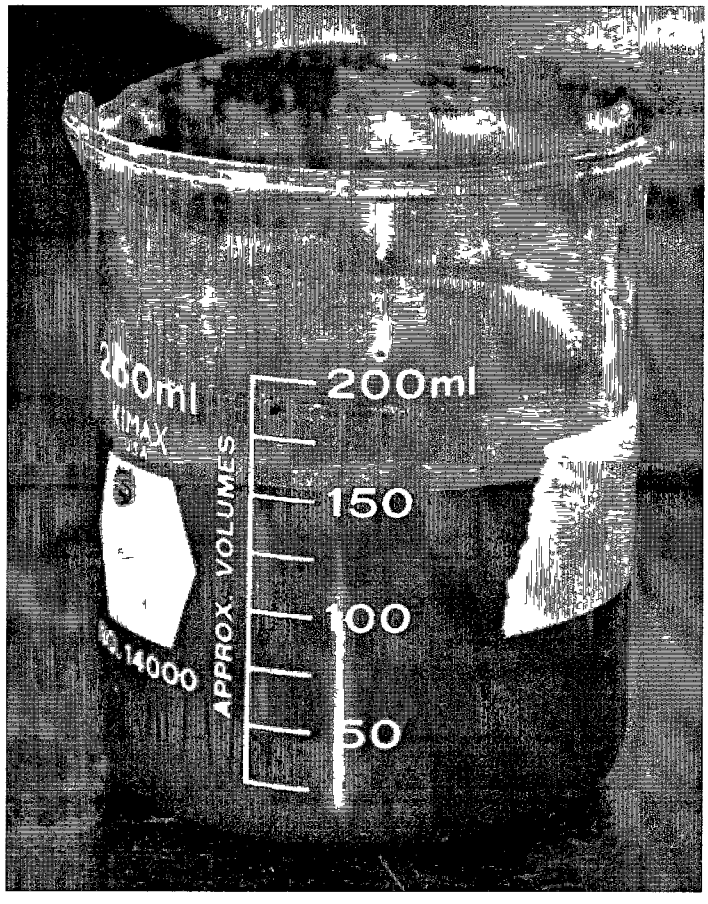


Figure 3.14 Ice-rich sample settled in a beaker. The supernatant water and saturated sediment volumes are used to calculate excess-ice content.

(4) $EE = [(W_v * 1.09)/(S_v + (W_v * 1.09))] * 100$

where EE is the excess-ice content (%), W_v is the volume of supernatant water, S_v is the volume of saturated sediment, and 1.09 represents the expansion of frozen water.

3.7.3 Organic matter content

To examine the influence of organic matter content on the accumulation of near-surface ground ice, subsamples of soil from the top of permafrost were analyzed using the loss-on-ignition method (Sheldrick, 1984). One 10 cm section from the top 30 cm of permafrost was retained from each drill site. Since basin soils appeared homogenous at the top of permafrost during drilling, this sample was considered as representative of the organic matter content for this zone. Subsamples of 1 cm^3 were measured and placed in crucibles, and then oven-dried at 105°C overnight. These samples were cooled in a desiccator and weighed. The organic matter was removed by heating the samples to 550°C in a muffle furnace. The crucibles were again allowed to cool in a desiccator and reweighed, and the organic matter determined according to (Sheldrick, 1984):

(5) $LOI = ((S_{OD} - S_i) / S_{OD}) * 100$

where LOI is the % loss-on-ignition, S_{OD} is the mass of sediment after oven drying, and S_i is the sediment mass after ignition at 550°C.

3.7.4 Particle size

Soil particle size distribution was determined using the laser diffraction (LD) method (e.g. Eshel et al., 2004). The main advantages of this technique include a short analysis time, repeatable results, and the small amount of sample required. In addition, the output of the analysis provides particle size fractions for a large number of sizes over

the measurement range. Therefore, a continuous particle size distribution can be obtained (Eshel et al., 2004).

Soil subsamples were taken from the same 10 cm sections that were used for organic matter content determination. When extracting the cores, soil at the top of permafrost in the basin was generally homogenous, so the 10 cm section from the top 30 cm of permafrost represented soil conditions of near-surface permafrost. To prepare the soil samples, each section was crushed and homogenised using a mortar and pestle. Following this, 1 cm^3 subsamples were measured and subsequently put through a 2 mm sieve to ensure no coarse particles were present that could damage the laser diffractor. All soil passed through the sieve, although a few pieces of intact organic matter did not pass through the sieve and were removed. The samples were then placed in beakers.

Organic matter was digested by adding hydrogen peroxide to the soil sample and heating the beakers on hot plates to speed the reaction (Sheldrick, 1984). Following the organic digestion, samples were transferred to centrifuge tubes. Distilled water was added to the samples, which were then centrifuged. This process was repeated three times in order to rinse any remaining hydrogen peroxide from the sample.

Next, samples were transferred into crucibles and the water evaporated in a drying oven. Calgon was added to the samples to prevent flocculation (Konert and Vandenberghe, 1997). The particle-size distribution was obtained using a Beckman Coulter LS 13 320 laser diffraction particle size analyzer in Carleton University's Department of Earth Sciences (see www.beckmancoulter.com). The instrument measures particle sizes between 0.375 and 2000 μm and reports the volume percentage of 93 size

bands within this range of measurement. The output also provides descriptive statistics of the particle-size distribution.

Limitations of this method have been documented in a few studies. The greatest limitation of LD is likely the common underestimation of the clay fraction (e.g. Konert et al., 1997; Beuselinck et al.; Campbell, 2003) compared to the pipette method (see Sheldrick, 1984). However, this is only an issue when directly comparing data obtained by the two methods. The results obtained by LD can be processed to obtain results comparable to the pipette method, as demonstrated by Konert et al. (1997). In this study, all sites were analyzed using LD, so there is no problem of comparison between the two methods.

A possible source of error, though likely minimal, may be a small loss of sediment when transferring the samples between the beakers, centrifuge tubes, and crucibles. Some amount of sediment may have been lost during these preparatory steps. However, each container was rinsed thoroughly with deionised water in order to minimize sediment loss.

The LD method was chosen because of the mentioned advantages, and because LD will likely become more widely used in the future for particle-size distribution analysis (Eshel et al., 2004).

4. RESULTS

4.1 Introduction

Ice has formed in the sediments of Illisarvik since lake drainage as permafrost has aggraded into the exposed sub-lake talik. The first section of this chapter examines the active-layer history in the basin, tundra and terrace, because active-layer dynamics are inherently linked to the growth and thaw of near-surface ground ice. Effects of climate and snow accumulation on basin, tundra and terrace active-layer depths are assessed. Following this, the amount and distribution of ice-rich, near-surface permafrost in the basin, tundra, and lake terraces are presented. Relations between ice enrichment and its potential controlling factors are subsequently described. Finally, comparisons of moisture content profiles from data collected in 1979 and 2010 are presented.

4.2 Basin active-layer history

The active-layer history from 1979 to 1999 was examined using data collected from active-layer transects along the major and minor axes of the lake (Mackay and Burn, 2002a), supplemented by data collected by Dr. C.R. Burn between 2000-2010. The active-layer transect benchmarks were grouped into three units in this study based on vegetation and site characteristics (Figure 4.1): (1) wet willow areas (WW) (Figure 4.2), (2) drier grassy areas near the lake basin centre (BC) (Figure 4.2), and (3) sandy areas near lake margins/areas with little vegetation (LM) (Figure 4.3). Of 35 active-layer benchmarks in the lake basin, seven were in WW, thirteen in BC, and six in LM (Figure 4.1). Of the nine remaining sites, eight near South Pond were not included in the analysis

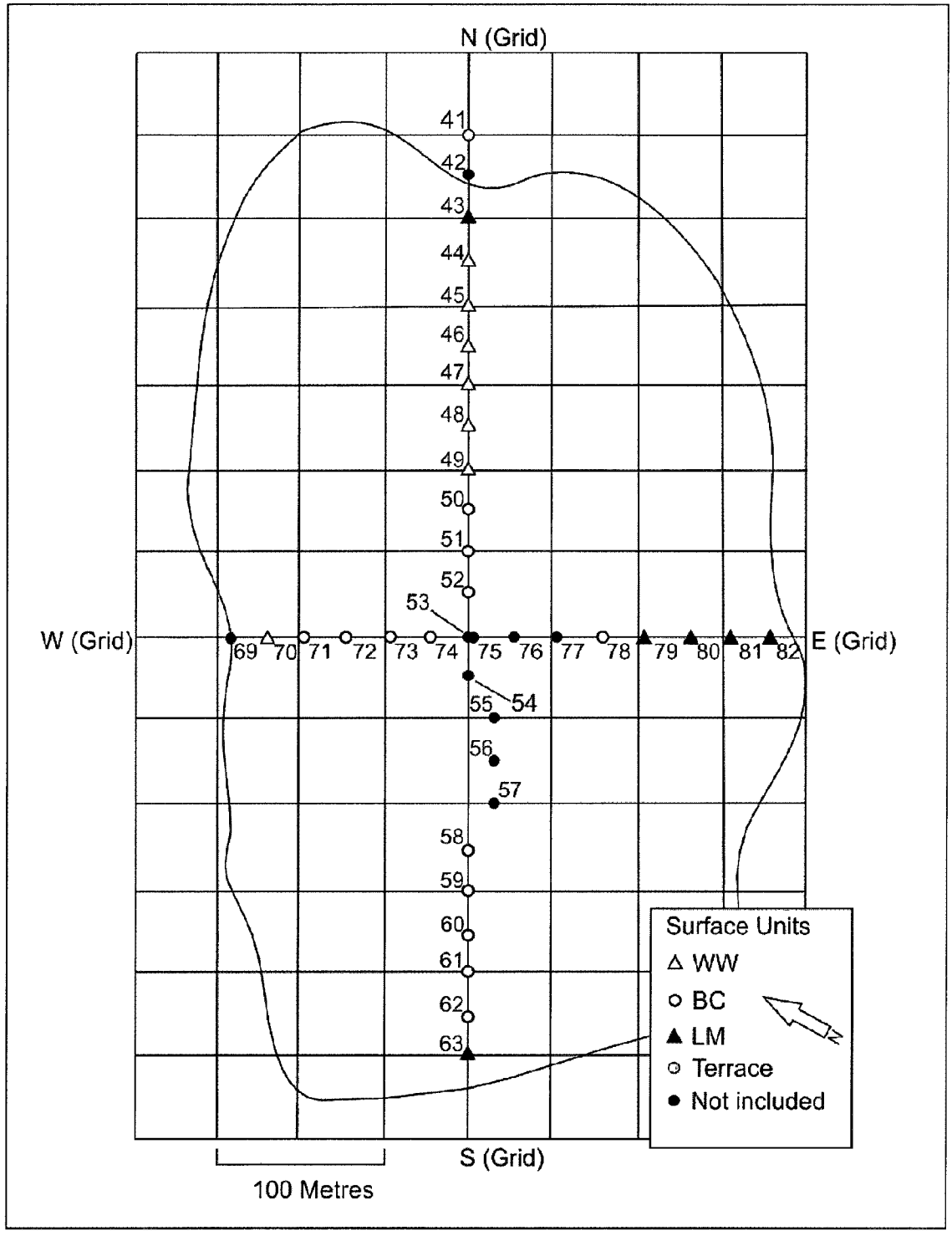


Figure 4.1 Active layer benchmark classifications (WW = wet willows; BC = basin centre; LM = lake margins/little vegetation.

a)

b)

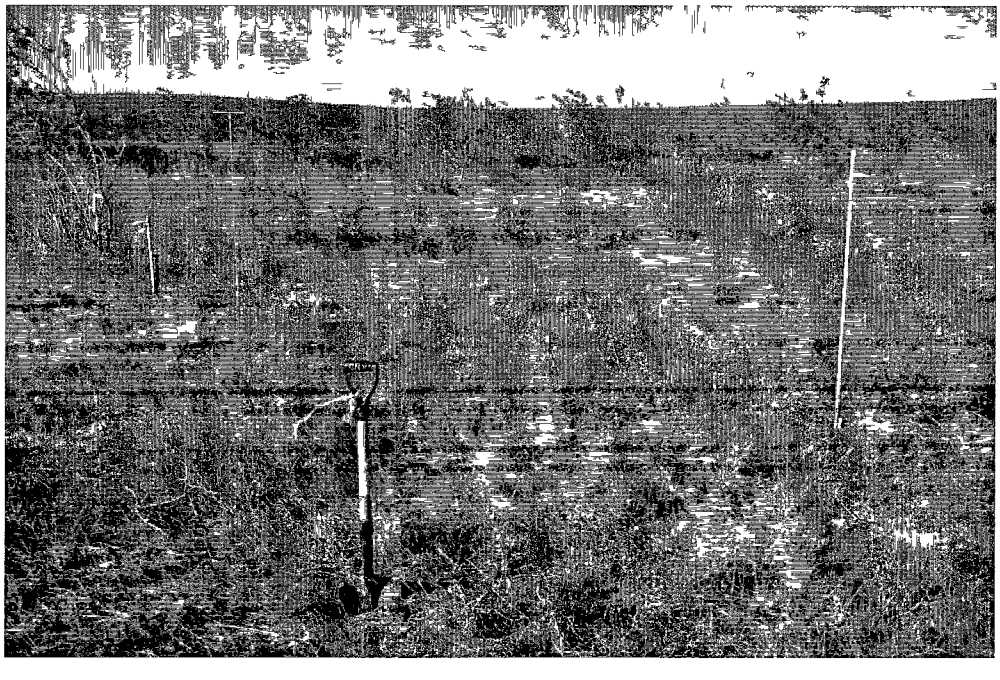


Figure 4.2 (a) Wet willow (WW) drill site and (b) basin centre (BC) drill site.

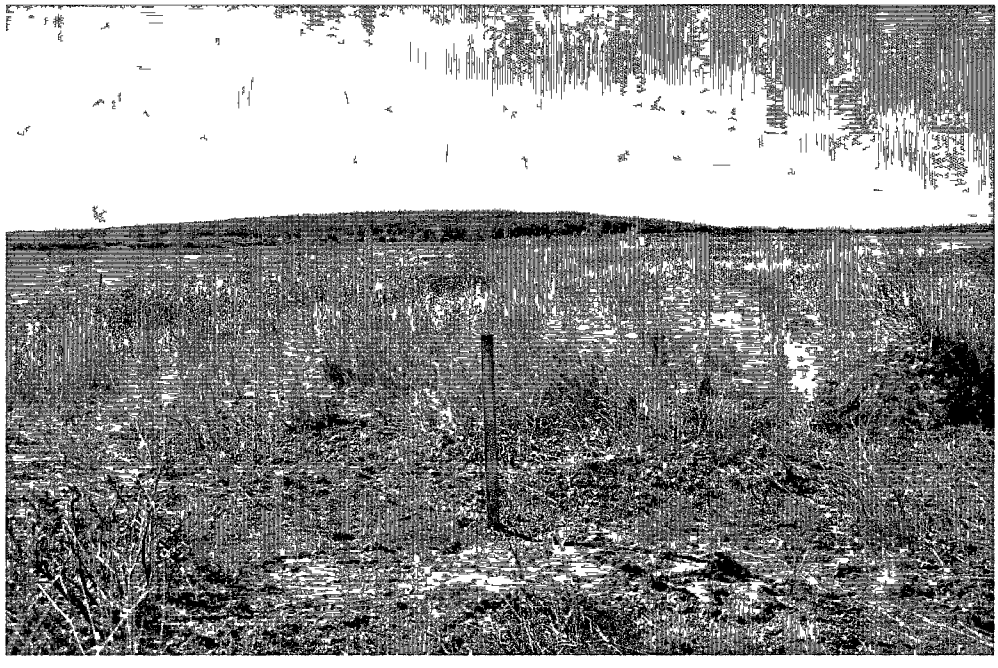


Figure 4.3 Basin drill site in an area with little vegetation (LM).

because flooding during some years affected the collection of active-layer depths and the records from these locations are not complete (sites 75-77 and 53-57). Benchmark 69 was not included in the analysis due to its unique position at the interface between steep lake shore and basin bottom. The benchmarks assigned to each class are in Table 4-1.

The active-layer thicknesses since 1979 for each basin unit are shown in Figure 4.4. Autocorrelation analysis was conducted using the mean active-layer depths for each of the three surface classes, to determine if conditions and events in previous years influenced active-layer thickness as suggested by Miller et al. (1998) and Shur et al. (2005). The autocorrelation coefficients were determined over 16 lags (1 year per lag) and examined at the 95% confidence level. At all three basin units, there were no significant autocorrelations at lag 1, suggesting that the yearly mean active-layer thicknesses are not influenced by conditions in previous years.

To examine the active-layer history of each class, the Mann-Whitney U test was used to test decadal-scale differences in active-layer depths. The data for the WW, BC and LM classes were split into three periods, 1979-1989, 1990-1999, and 2000-2010.

At WW sites, there was no difference in mean active-layer depth between 1979-1989 ($\bar{x}$ = 67, s = 4.8) and 1990-1999 ($\bar{x}$ = 67 cm, s = 3.1 cm); U(21) = 49.5 , p = 0.698 (Figure 4.4a). However, between 1990-1999 and 2000-2010 ($\bar{x}$ = 60 cm, s = 4.5 cm) there was a significant difference in active-layer depths; U(21) = 16 , $p < .01$. The average depth in the 2000s was 7 cm less than in the 1990s.

Table 4-1 Basin active-layer benchmark classifications

Basin Class	Benchmarks
Wet willow	44, 45, 46, 47, 48, 49, 70
Basin centre	50, 51, 52, 58, 59, 60, 61, 62, 71, 72, 73, 74, 78
Margins/little vegetation	43, 63, 79, 80, 81, 82

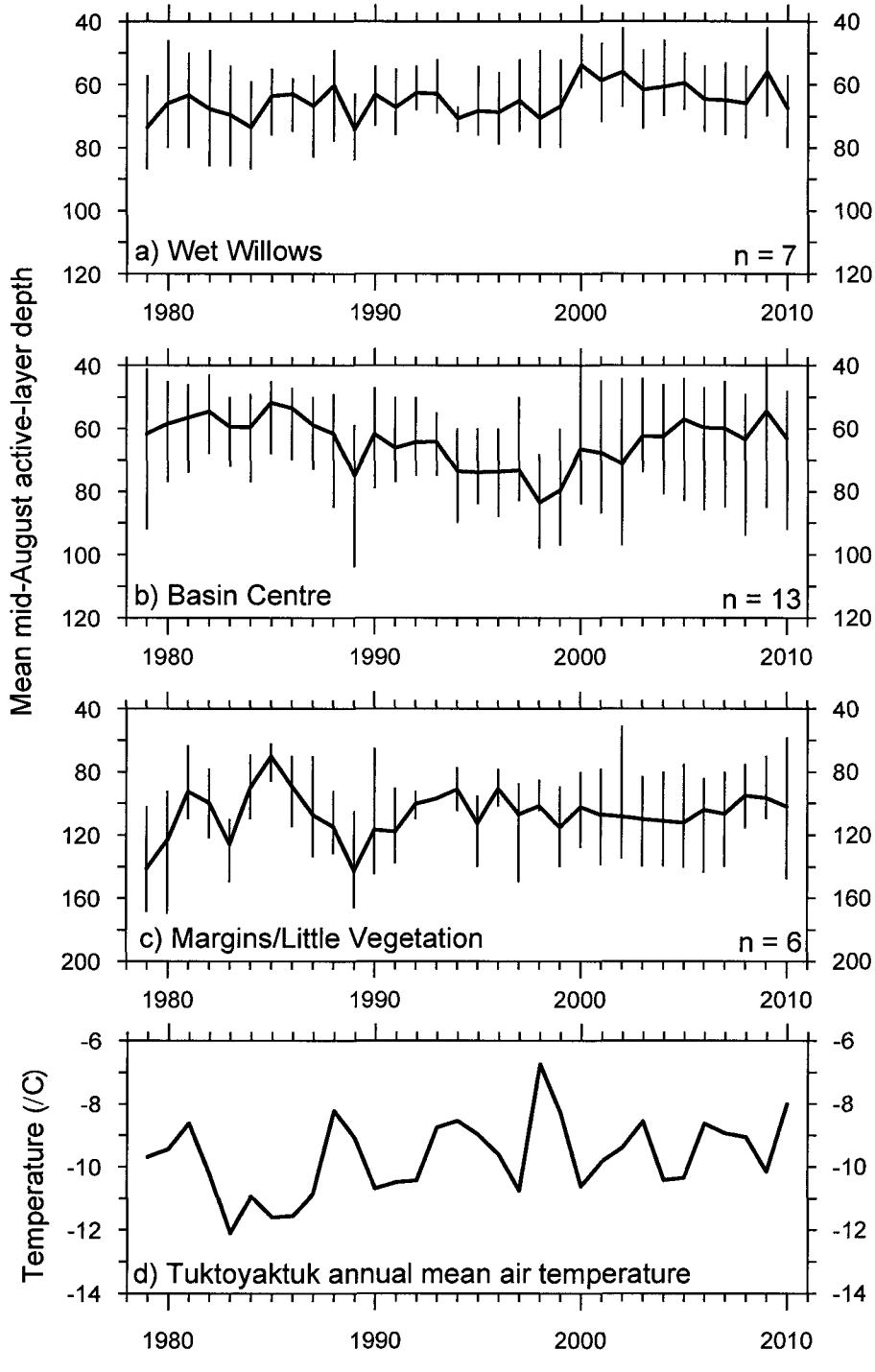


Figure 4.4 Average active-layer depths for (a) wet willow sites, (b) dryer sites near the lake centre, (c) margin sites/sites with little vegetation, and (d) annual mean air temperature at Tuktoyaktuk A. Vertical lines indicate the range of active layers of all sites within the basin unit.

BC sites had significantly different mean active-layer depths between 1979-1989 ($\bar{x}$ = 59 cm, s = 3.4 cm) and 1990-1999 ($\bar{x}$ = 71 cm, s = 7.1 cm); U(21) = 8.0 , $p < 0.01$ (Figure 4.4b). The average active-layer depth in the 1990s was 12 cm greater than in the 1980s at these sites. During the 2000s, the average active-layer depth was 8 cm thinner than in the 1990s, which was a significant difference [2000-2010 depths ($\bar{x}$ = 63 cm, s = 4.8 cm)]; U(20) = 16.0 , $p < 0.01$.

LM sites saw no significant differences in average active-layer depths between 1979-1989 ($\bar{x}$ = 109 cm, s = 2.3 cm) and 1990-1999 ($\bar{x}$ = 105 cm, s = 1.0 cm); U(21) = 52.0 , p = 0.823, or between 1990-1999 and 2000-2010 ($\bar{x}$ = 105 cm, s = 5.6 cm); U(21) = 54 , p = 0.944 (Figure 4.4c).

In summary, active-layer thicknesses were stable between the 1980s and 1990s, and then thinned in the 2000s at WW sites. Mean active-layer thicknesses in BC increased between the 1980s and 1990s, and subsequently thinned in the 2000s. In LM, mean active layers have remained stable in the three decades since drainage, though the variability has decreased. These results are different from the tundra active-layer history presented by Mackay and Burn (2002a), where active layers generally thickened between 1983-1999, but the basin setting is quite different than the tundra, and local conditions such as vegetation have been more variable and dynamic at basin sites.

4.2.1 Controls of basin active-layer history

4.2.1.1 Air temperature

To examine the variables responsible for active-layer variation in the basin, correlations between air temperature and snow depth and active-layer thickness were examined. Monthly average air temperatures from Tuktoyaktuk were averaged to obtain the annual mean air temperature (AMAT) and the mean temperature for each season [Sept-Nov (autumn), Dec-Feb (winter), Mar-May (spring), Jun-Aug (summer)] for all years since drainage. For this analysis, 'year' refers to the period from September to August, as the conditions during active-layer freeze back in the fall may influence the active-layer depth in the following summer, and conditions subsequent to the date of measurement will not be physically related to active-layer thickness.

Seasonal air temperatures were correlated with mean active-layer depths for each of the four basin class types addressed in the previous section. The correlations were examined for the periods 1979-1989, 1990-1999, and 2000-2010 (see Table 4-2).

Winter and spring air temperatures were not significantly correlated with active-layer depths in any decade. Autumn air temperatures were significantly correlated with active-layer thickness for 1990-1999 at WW and BC, which may suggest that temperature conditions during freeze-back have influenced active-layer depths the following summer during the 1990s. Annual mean air temperatures were also significantly correlated with active-layer depths at WW, and BC between 1990 and 1999. Though these correlations were statistically significant, the r^2 values were not always high, suggesting that other factors may be important for the resulting active-layer depths.

Table 4-2 Correlation coefficients and coefficients of determination for basin active-layer depths and air temperatures.

	Wet Willow		Basin Centre		Margins/Little Vegetation	
AMAT	r	r^2	r	r^2	r	r^2
1979-1989	-0.13	0.02	0.46	0.21	0.38	0.15
1990-1999	**0.62***	**0.38**	**0.74***	**0.55**	-0.29	0.09
2000-2010	**0.78***	**0.61**	0.04	0.00	-0.16	0.03
Sep-Nov						
1979-1989	-0.37	0.14	-0.29	0.08	0.06	0.00
1990-1999	**0.64***	**0.41**	**0.70***	**0.49**	-0.24	0.06
2000-2010	0.33	0.11	0.16	0.03	0.19	0.03
Dec-Feb						
1979-1989	-0.15	0.02	0.45	0.20	0.31	0.10
1990-1999	0.28	0.08	0.46	0.21	-0.24	0.06
2000-2010	0.45	0.20	-0.15	0.02	-0.38	0.14
Mar-May						
1979-1989	-0.23	0.05	0.20	0.04	-0.15	0.02
1990-1999	0.32	0.10	0.24	0.06	0.05	0.00
2000-2010	0.47	0.22	0.02	0.00	-0.06	0.00
June-Aug						
1979-1989	0.52	0.27	**0.90***	**0.81**	**0.87***	**0.75**
1990-1999	0.34	0.11	**0.55***	**0.30**	-0.38	0.15
2000-2010	**0.67***	**0.45**	0.13	0.02	-0.10	0.01

*Significant correlation coefficient at 95 % confidence level.

Perhaps the most interesting correlations are those with summer temperatures. From 1979-1989, active-layer depths were significantly and strongly correlated with summer temperature in BC and LM (Table 4-2). In the 1990s, the correlation coefficients were lower in magnitude and only significant for BC. The BC correlation coefficient decreased again in the 2000s, and was insignificant. The decrease in correlation coefficients suggests an increasing influence on active-layer depths by vegetation effects such as shading and insulation. In the first decade following drainage when vegetation cover was sparse or absent in much of the basin, the active layer likely responded mainly to summer temperatures, when the r^2 values were between 0.75 and 0.81. In these areas, active-layer development may be considered 'climate-driven' during the first decade following drainage (Shur and Jorgenson, 2007). The associations changed in the 1990s, where summer temperatures were only significantly correlated to active-layer depth in the basin centre, and to a lesser extent ($r^2 = 0.30$). It is suggested that during the 1990s, the establishing vegetation cover had significantly altered the summer energy balance at the ground surface, increasing shading and insulation, and these site-specific conditions dominated the influence of summer air temperatures. Because vegetation likely altered the surface and influenced the active layer, the thermal regime in the 1990s could be considered 'climate-driven, ecosystem modified' (Shur and Jorgenson, 2007).

Initial vegetation establishment was most rapid around North Pond, where the WW sites are located (Mackay and Burn, 2002b). Summer insulation and shading from vegetation may have been substantial even in the first decade following drainage, which may explain the weak correlation observed between summer air temperature at WW between 1979-1989 and 1990-1999. The permafrost was likely responding to these

modified surface conditions during the first decade and could therefore be considered 'climate-driven, ecosystem modified'.

4.2.1.2 Snow depths

The associations between late winter snow depths and active-layer depths at active-layer transect benchmarks were examined. Snow depths between 1980 and 2010 were used for this analysis, however depths were not available for 1990-1996 and 2006. The average snow depth for each year was calculated for sites in the three lake-bottom units (Figure 4.5).

Results from linear regressions suggest that snow depths in all classes have been increasing since 1980. The mean snow depth over the period of record was highest in WW (41 cm), and the increase in depth over time was significant at the 95% confidence level ($r^2 = 0.49$; $p < 0.01$). In the basin centre, the mean snow depth since 1980 was 29 cm, and the increase in depth over time was also statistically significant ($r^2 = 0.75$; $p < 0.01$). Finally, the mean snow depth since 1980 in LM was 38 cm, and the increase with time was statistically significant ($r^2 = 0.34$; $p < 0.01$). The increases in snow depths are likely due to progressive entrapment by vegetation, as snow depths were highest in WW, where the vegetation cover is highest and most widespread. Conversely, increasing snow depths may be a result of increased precipitation, but there is little evidence for this in the climate records at Tuktoyaktuk (Figure 3.3 and 4.5).

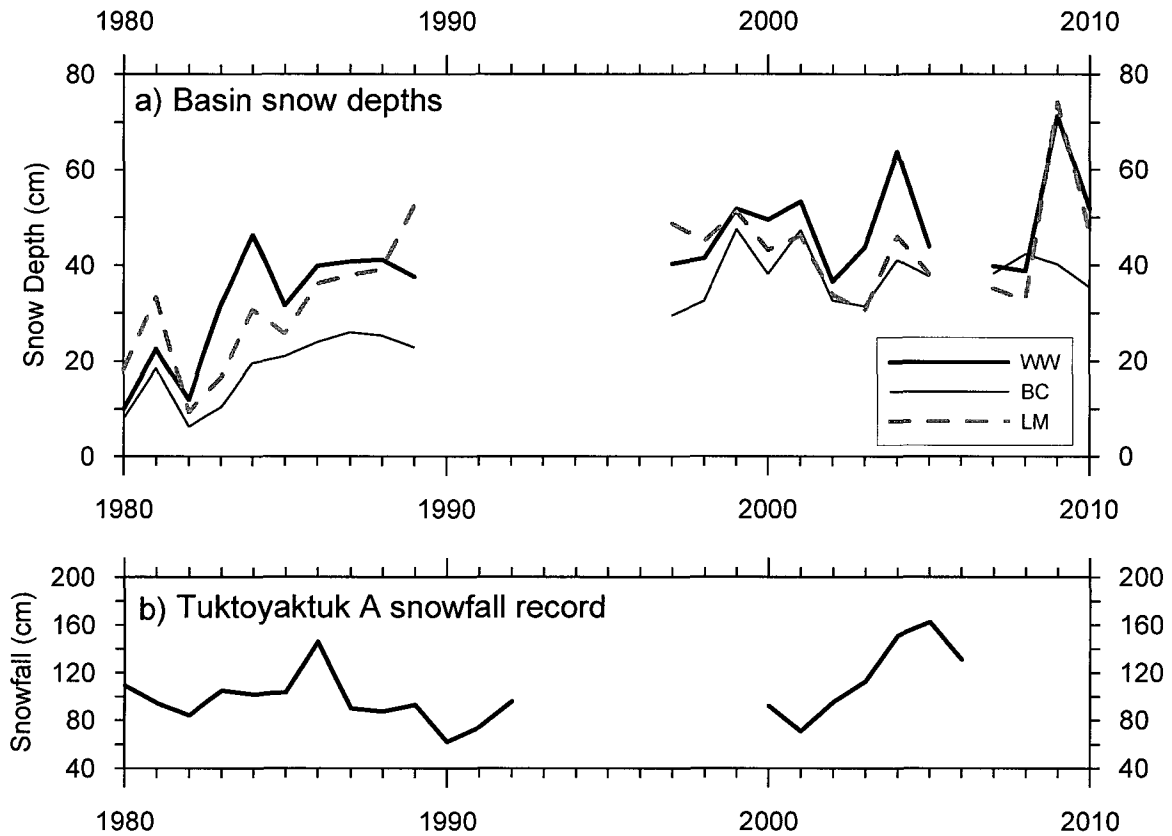


Figure 4.5 a) Mean lake basin snow depths at wet willow (WW), basin centre (BC), and lake margin/little vegetation sites (LM). Snow depths have been increasing in WW (r^2 = 0.49; $p < 0.01$), BC (r^2 = 0.751; $p < 0.01$), SP (r^2 = 0.456; $p < 0.01$) and LM (r^2 = 0.343; $p < 0.01$) and b) snowfall record from Tuktoyaktuk A weather station.

The snow depths for all available years were correlated with active-layer depths (Table 4-3). There were no significant correlations between active layers and snow depths in WW and LM. A significant correlation was found at BC sites ($r = 0.40$; $p = 0.03$). These results suggest that snow depth may not be strongly related to active-layer depths, but may be of some importance in certain areas of the basin.

4.2.1.3 Multiple regression

To examine the combined effects of seasonal temperatures and snow depths on basin active-layer depths, multiple regression was employed for each basin unit. The regressions were performed using the stepwise method, with snow depth and autumn, winter, spring, and summer temperatures as possible independent variables, and the active layer depth as the dependent variable. Summary results of the regressions are presented in Table 4-4. Snow depths significantly influenced the models for WW and BC. Summer air temperature was the only seasonal air temperature that significantly influenced the models in all basin units. The lowest r^2 was 0.313 for LM, while the highest was observed at BC sites (0.613). The r^2 for WW areas was 0.391. These regression results, along with the correlations presented in sections 4.2.1.1 and 4.2.1.2 suggest that the active layers in the lake basin are complex systems that respond to a number of variables, and that one or more influential variables were not examined in these analyses. Soil moisture, latent heat requirements to melt ground ice, and shading and surface organic matter accumulation have not been measured since drainage and these may be contributing to basin active-layer depth variations.

Table 4-3 Correlation coefficients and coefficients of determination between basin active-layer depths and snow depths. Snow depths were not available for 1990-1996 and 2006.

	Wet Willow		Basin Centre		Margins/Little Vegetation	
	r	r^2	r	r^2	r	r^2
1980-2010	-0.35	0.12	**0.40***	**0.16***	0.09	0.01

*Significant correlation at the 95% confidence level.

Table 4-4 Summary results of stepwise multiple regressions for basin, tundra, and terrace classes. Fall, winter, spring, and summer refer to the mean seasonal temperatures.

	Independent Variables	**r**	**r^2**	**Std. Error**
Basin				
WW	Summer, Snow depths	0.625	0.391	4.388
BC	Summer, Snow depths	0.786	0.613	5.439
LM	Summer	0.559	0.313	12.287
Tundra				
Hillslope	Summer	0.743	0.552	2.880
Hilltop	Summer	0.598	0.357	4.606
Terrace				
Benchmark 41	Fall, winter, spring, summer, Snow depths	0.303	0.092	6.192

4.3 Tundra active-layer history

Active-layer depths from twelve benchmarks along the tundra transect have been collected since 1983 (Figure 4.6) (see Mackay and Burn, 2002a). The tundra transect active-layer benchmarks were grouped into four classes based on terrain type: (1) lowland sites (sites 1-5), (2) Site 6 in the trough below the slope, (3) hillslope sites (sites 7-9), and (4) hilltop sites (sites 10-12). Lowland sites and Site 6 were not included in this analysis because tundra drill sites were not located in these terrain types. The active-layer depths for hillslope and hilltop classes are presented in Figure 4.7.

Autocorrelation analysis of active layer depths was conducted for hillslope and hilltop tundra classes in the same manner as basin classes. There were no significant autocorrelations.

The Mann-Whitney U test was used to test decadal scale differences in active-layer depths at hillslope and hilltop sites. The data were split into three periods, 1983-1989, 1990-1999, and 2000-2010.

At hillslope sites, there was no significant difference in mean active-layer depths between 1983-1989 ($\bar{x}$ = 46 cm, s = 4.1 cm) and 1990-1999 ($\bar{x}$ = 48 cm, s = 4.0 cm); U(17) = 28.5 , p = 0.525. No significant difference existed in active layers between 1990-1999 and 2000-2010 ($\bar{x}$ = 47 cm, s = 3.7 cm); U(21) = 43.0 , p = 0.397.

For hilltop sites, there was no significant difference in mean active-layer depths at hilltop sites between 1983-1989 ($\bar{x}$ = 44 cm, s = 4.4 cm) and 1990-1999 ($\bar{x}$ = 48 cm, s = 6.2 cm); U(17) = 18.0 , p = 0.097. Active layer depths did not change significantly between 1990-2000 and 2000-2010 ($\bar{x}$ = 49 cm, s = 4.7 cm); U(21) = 44.5 , p = 0.470.

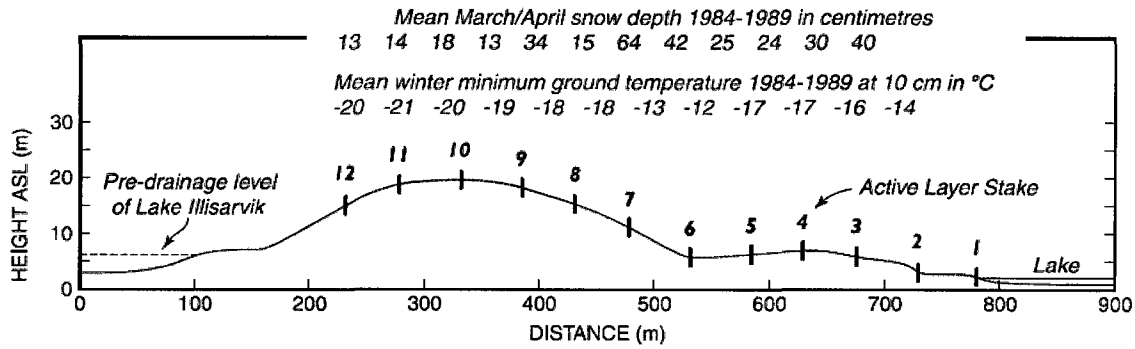


Figure 4.6 Tundra active-layer transect (Mackay and Burn, 2002a, Fig. 6).

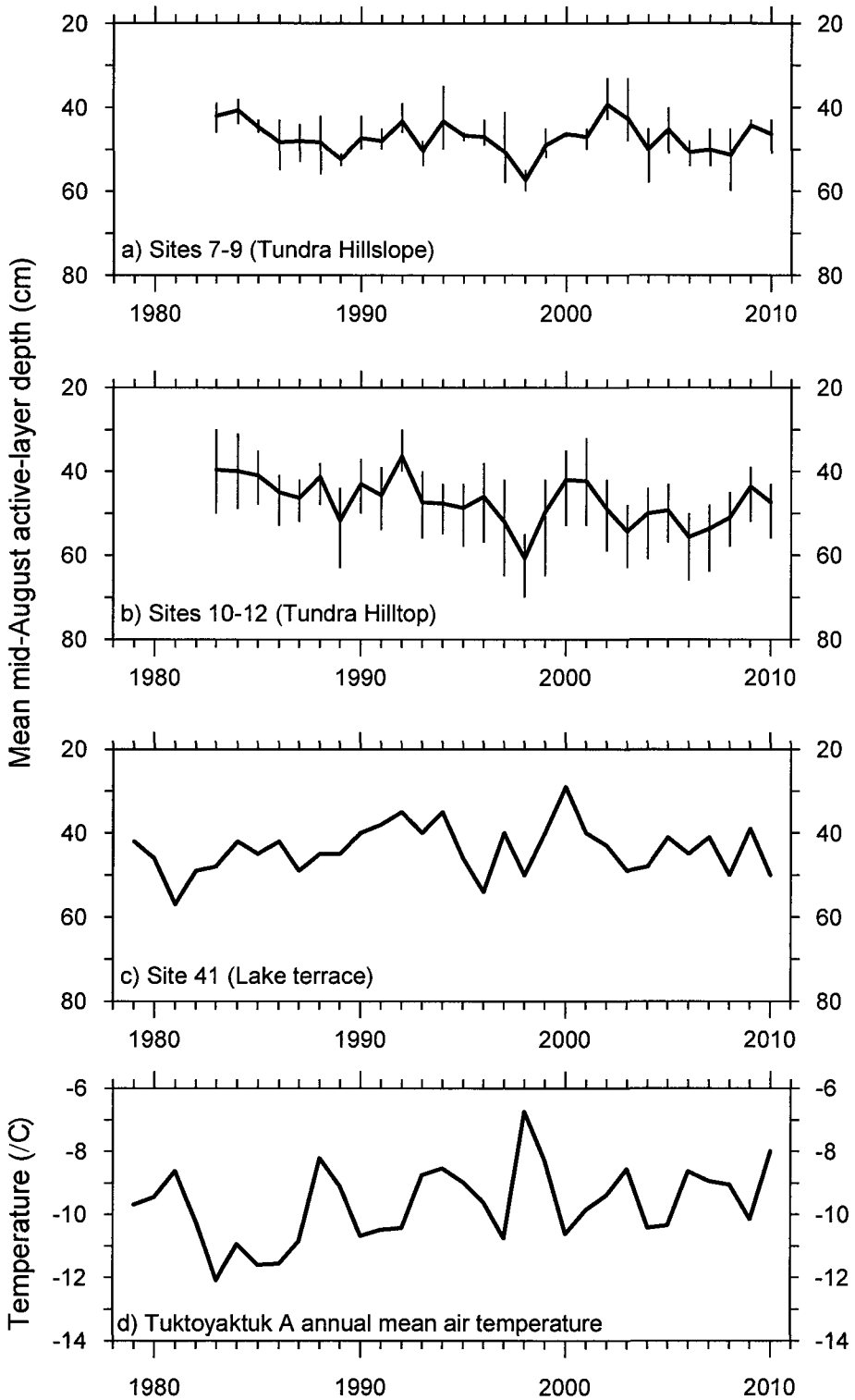


Figure 4.7 Active layer thicknesses at (a) hillslope and (b) hilltop sites along the 550 m tundra active-layer transect, (c) at active-layer stake Site 41 on the lake basin terrace, and (d) annual mean air temperature at Tuktoyaktuk A.

Despite the lack of significant differences between consecutive decades, there has been a gradual increase in active layer depths at hilltop sites (Figure 4.7). Linear regression between hilltop active-layer depths and time indicates that this trend is significant at the 95% confidence level ($r^2 = 0.26$, $p < 0.01$).

4.3.1 Controls of tundra active-layer thickness

4.3.1.1 Air temperature

Correlations between air temperatures and tundra hillslope and hilltop active-layer depths were examined in the same manner as basin classes. The periods examined were 1983-1989, 1990-1999 and 2000-2010. Active-layer depths at hillslope and hilltop sites in the 1980s were very strongly associated with winter temperatures (hillslope $r^2 = 0.86$; hilltop $r^2 = 0.63$) (Table 4-5). This is likely due to very cold winters in the early 1980s which resulted in shallow active layers, and the warm winter and summer of 1989 which resulted in deep thaw. Several other significant but weaker correlations existed between the autumn, spring, summer, and annual mean air temperatures, suggesting that air temperatures throughout the year may influence tundra active-layer depths.

4.3.1.2 Snow depths

The relations between late winter snow depths and active-layer depths in the tundra were examined. Snow depths between 1984 and 2010 were used for this analysis, however depths for 1990-1996 were not available. The average snow depths for each year were calculated for the hillslope and hilltop sites (Figure 4.8a).

Table 4-5 Correlation coefficients and coefficients of determination between tundra active-layer depths and air temperatures.

	Hillslope (Sites 7-9)		Hilltop (Sites 10-12)		Terrace	
AMAT	r	r^2	r	r^2	r	r^2
1983-1989[1]	0.61	0.38	0.38	0.15	0.29	0.08
1990-1999	0.53	0.28	**0.70***	**0.49***	0.38	0.14
2000-2010	0.07	0.00	**0.58***	**0.34***	**0.66***	**0.43***
1983-2010[1]	**0.43***	**0.18***	**0.64***	**0.41***	0.26	0.07
Sep-Nov						
1983-1989[1]	-0.06	0.00	-0.44	0.19	0.11	0.01
1990-1999	0.06	0.00	0.31	0.10	0.27	0.07
2000-2010	-0.08	0.01	**0.54***	**0.29***	0.54	0.29
1983-2010[1]	0.01	0.00	0.37	0.14	0.21	0.05
Dec-Feb						
1983-1989[1]	**0.93***	**0.86***	**0.79***	**0.63***	0.31	0.10
1990-1999	0.22	0.05	0.39	0.15	0.14	0.02
2000-2010	0.15	0.02	0.28	0.08	0.10	0.01
1983-2010[1]	0.36	0.13	**0.47***	**0.22***	0.18	0.03
Mar-May						
1983-1989[1]	0.18	0.03	-0.04	0.00	0.33	0.11
1990-1999	**0.63***	**0.40***	0.52	0.27	0.39	0.15
2000-2010	-0.37	0.14	0.24	0.06	0.50	0.25
1983-2010[1]	0.26	0.07	0.37	0.13	0.15	0.02
June-Aug						
1983-1989[1]	0.42	0.17	0.45	0.20	-0.07	0.01
1990-1999	0.50	0.25	**0.67***	**0.44***	0.14	0.02
2000-2010	**0.72***	**0.52***	0.39	0.16	0.47	0.22
1983-2010[1]	**0.55***	**0.30***	**0.47***	**0.22***	0.08	0.01

[1] For the lake terrace, the earliest year used was 1979.
*Significant correlation coefficient at 95 % confidence level.

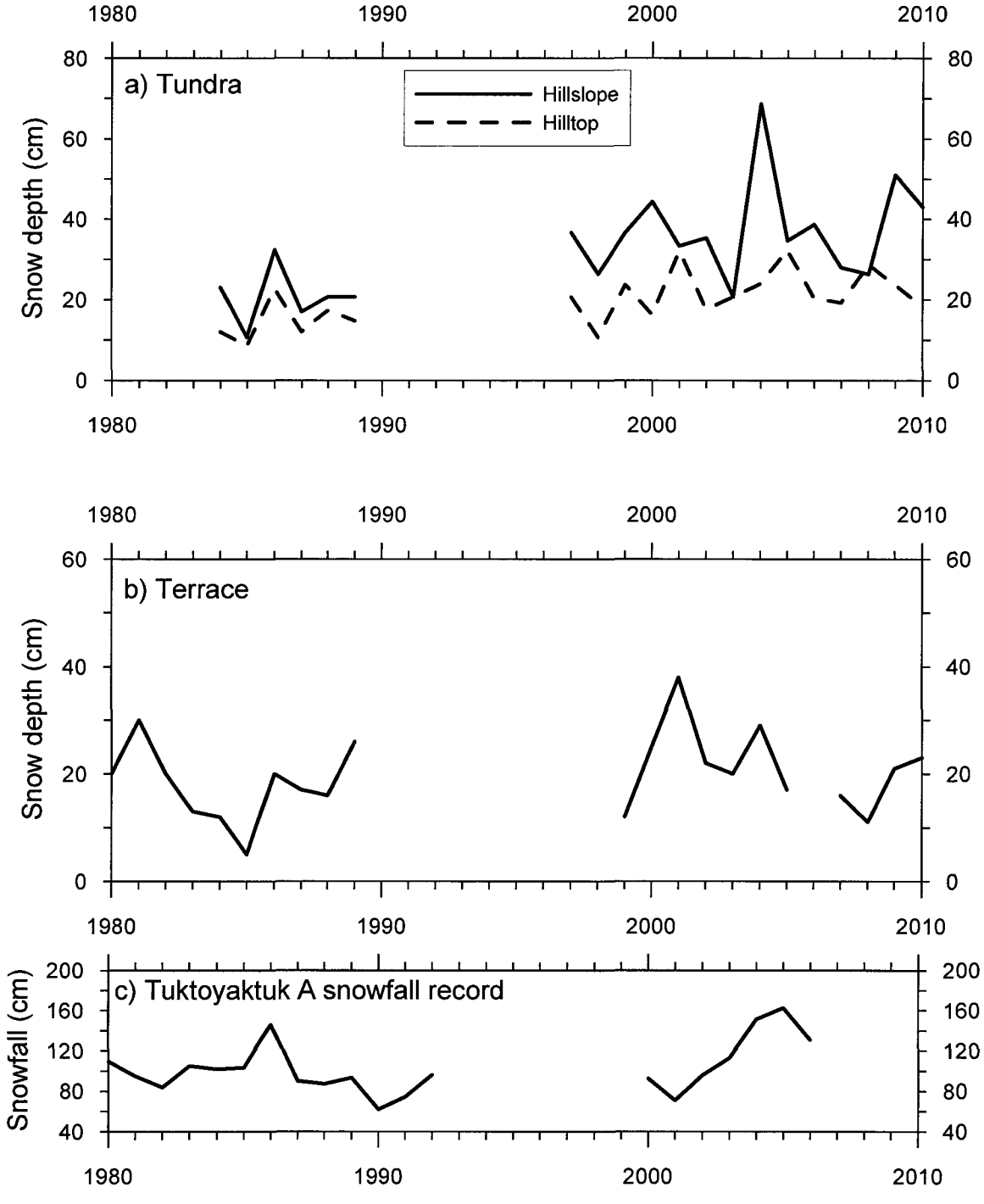


Figure 4.8 Snow depths for the (a) tundra hillslope and hilltop, (b) the lake basin terrace at Benchmark 41, and snowfall at (c) the Tuktoyaktuk A weather station . Linear regression indicated that snow depths increased on the hillslope ($r^2 = 0.35$; $p < 0.01$) and on the hilltop ($r^2 = 0.33$; $p < 0.01$), but there was no significant trend on the lake basin terrace ($r^2 = 0.03$; $p = 0.461$) or in the Tuktoyaktuk snowfall record ($r^2 = 0.13$; $p = 0.124$).

Results from linear regressions suggest that snow depths on the tundra hillslope and hilltop have been increasing since 1984. For sites 7-9 on the hillslope, the mean snow depth since 1984 was 32 cm, and the increase with depth over time was significant at the 95% confidence level ($r^2 = 0.35$; $p < 0.01$). On the hilltop, at sites 10-12, the mean snow depth since 1984 was 20 cm, and snow depths have also been increasing significantly over that period ($r^2 = 0.33$; $p < 0.01$). Snow depths on the tundra may be increasing as a result of vegetation growth.

The snow depths for all available years were correlated with active-layer depths (Table 4-6). There were no significant correlations between active layers and snow depths. These results suggest that the recorded snow depths may not have sufficient variation to be important to tundra active-layer development.

4.3.1.3 Multiple regressions

Multiple regression was employed for each tundra class. The regressions were performed using the stepwise method, with snow depth and fall, winter, spring, and summer temperatures as possible independent variables, and the active layer depth as the dependent variable. Summary results of the regressions are in Table 4-4. Summer temperature was the only independent variable that significantly influenced the model. The coefficients of determination were similar to those observed in the lake basin for the hillslope ($r^2 = 0.55$) and hilltop ($r^2 = 0.35$), suggesting that, like in the lake basin, tundra active layers also respond to other variables. Ground ice, soil moisture, surface and sub-surface organic matter accumulation may be variables that influence tundra active layers.

Table 4-6 Correlation coefficients and coefficients of determination for tundra and terrace active-layer depths and snow depths.

	Hillslope		Hilltop		Terrace	
	r	r^2	r	r^2	r	r^2
1984-2010	0.03	0.00	0.04	0.00	-0.03	0.00

4.4 Terrace active-layer history

The active-layer history on former lake terraces was examined using data collected between 1979 and 2010 from the active-layer transect along the major basin axis. Benchmark 41 was used as a representative terrace site. Benchmark 42, though also on the terrace, is near the shoreline and has much deeper active-layer depths than the sites drilled on terraces, so it was not included. Terrace active-layer thicknesses are shown in Figure 4.7.

Autocorrelation coefficients were determined as in section 4.2, and none were significant. Decadal differences in active-layer thicknesses were examined for Benchmark 41 using the Mann-Whitney U test and methodology described in section 4.2. There were no significant differences in active-layer thicknesses between decades.

4.4.1 Controls of terrace active layers

Correlations between air temperatures and terrace active-layer depths at Benchmark 41 were examined in the same manner as basin classes (Table 4-5). The only significant correlation occurred between the annual mean air temperature and active layers from 2000-2010. The lack of association between air temperatures and active layers compared to basin and tundra sites may be a result of the thick layer of peat found on terraces, which, due to its low thermal conductivity when dry, may insulate the active-layer during the summer. Also, the heat required to melt ice in the peat, and the ability of the peat to retain water once thawed likely reduces its sensitivity to temperature due to the lower thermal conductivity in the liquid phase.

The relations between late winter snow depths and active-layer depths on the terrace were examined. Snow depths between 1980 and 2010 were used for this analysis, however, depths for 1990-1996 and 2006 were not available. The snow depths used were from Benchmark 41, located on terrace along the major axis active-layer transect (Figure 4.8b). There was no significant correlation between active layers and snow depths (Table 4-6). These results suggest that snow depth may not be an important factor to the active-layer development on the terrace.

Results from linear regression indicate that snow depths on the tundra have not been changing since 1980. At Benchmark 41, the mean snow depth since 1980 was 20 cm, and there was no trend in snow depths over the period of record ($r^2 = 0.026$; $p = 0.461$). The vegetation at benchmark 41 on the lake terrace has remained essentially unchanged over the period of record. The absence of increasing snow depths on the terrace suggest that the increases observed in the basin and tundra are likely a result of augmented snow entrapment from growing vegetation rather than an increased snow supply.

4.4.1.1 Multiple regession

Multiple regression was employed for terrace active layer depths. The regression was performed using the stepwise method, with snow depth and fall, winter, spring, and summer temperatures as possible independent variables, and the active layer depth as the dependent variable. None of the variables significantly predicted active-layer depths. When all independent variables were entered in the regression, the resulting r^2 was 0.092. This result suggests that air temperatures and snow depths do not influence the terrace active layer. As previously mentioned, the terrace is covered in a thick layer of peat, and

these unique surface conditions may explain the insensitivity to air temperatures and snow of the terrace active layer.

4.5 Excess-ice accumulation at the top of permafrost

4.5.1 Lake basin

The accumulation of ice in near-surface permafrost was examined at 26 sites in the drained lake basin. Ice enrichment of the upper 100 cm of permafrost was highly variable in the basin (Figure 4.9). Seventeen of the 26 sites were enriched with excess ice, and the other 9 sites were ice-poor. The majority of ice-rich sites occurred in wet ground near North Pond. Sites along the southern lake margins were almost exclusively ice-poor.

The maximum depth of excess ice was 60 cm below the 2010 permafrost table (Figure 4.10b). Mackay and Burn (2002a) reported that aggradational ice had grown in the basin to depths of at least 40 cm since drainage, and so the results from this study confirm that ice has accumulated at depths beyond those previously reported. For ice-rich sites, the average excess-ice content of the upper 50 cm of permafrost was 8%. The maximum excess-ice content of the upper 50 cm was 19% at Site 7. The mean accumulation in the top of 100 cm of permafrost corresponds to 3.6 cm of ice. The maximum amount of ice development was 9.7 cm at Site 7. Ice accumulation at each site is presented in Table 4-7.

4.5.1.1 Wet willows

Ice content at drill sites were also examined based on the classification described in section 4.2. Eight drill sites were located in WW (Table 4-7). These sites were in very soggy ground, commonly near standing water and *Equisetum* spp. plants. The base of the

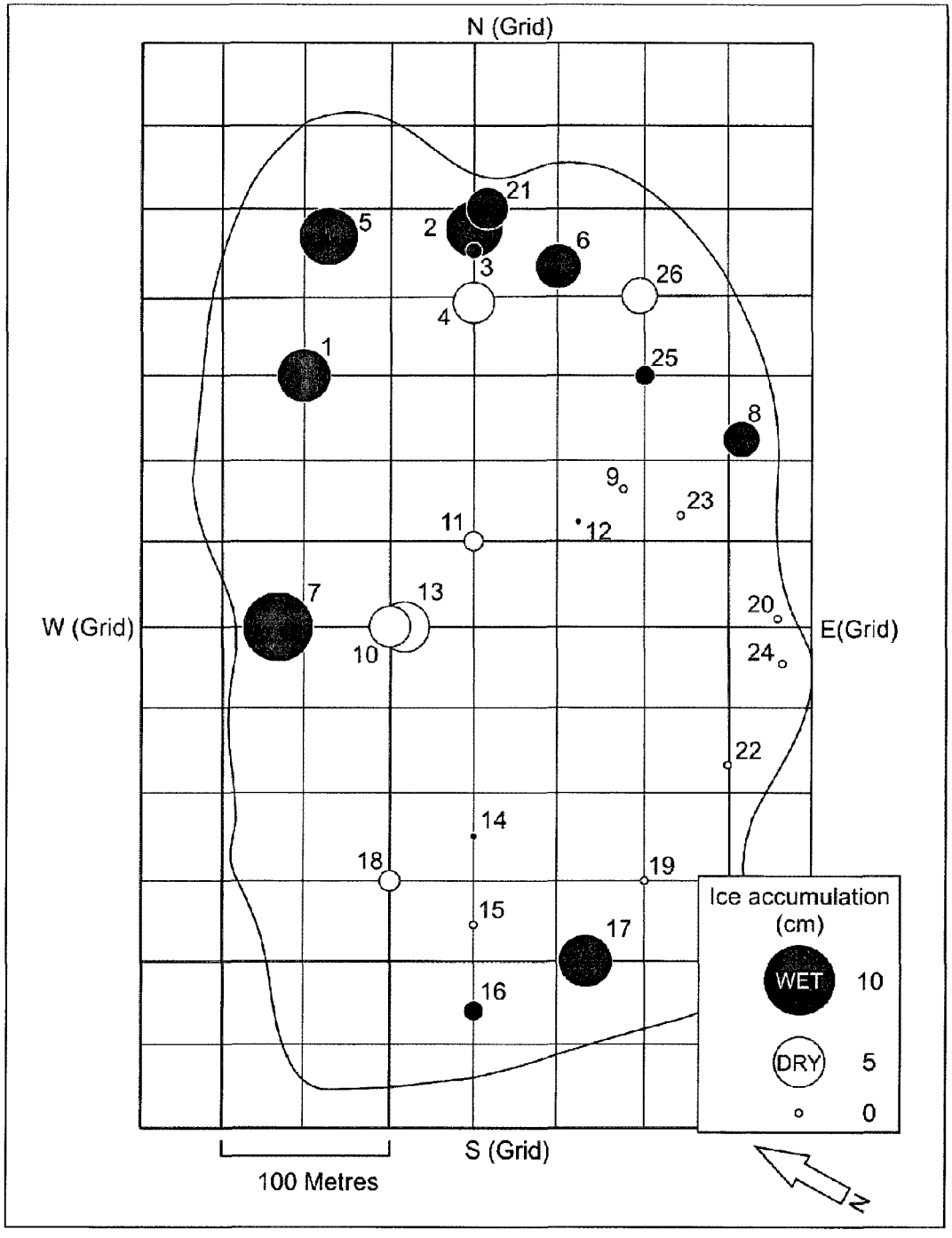


Figure 4.9 Proportional symbol map of ice accumulation of the top 100 cm of permafrost at basin sites. The radius of the circle is proportional to the ice accumulation.

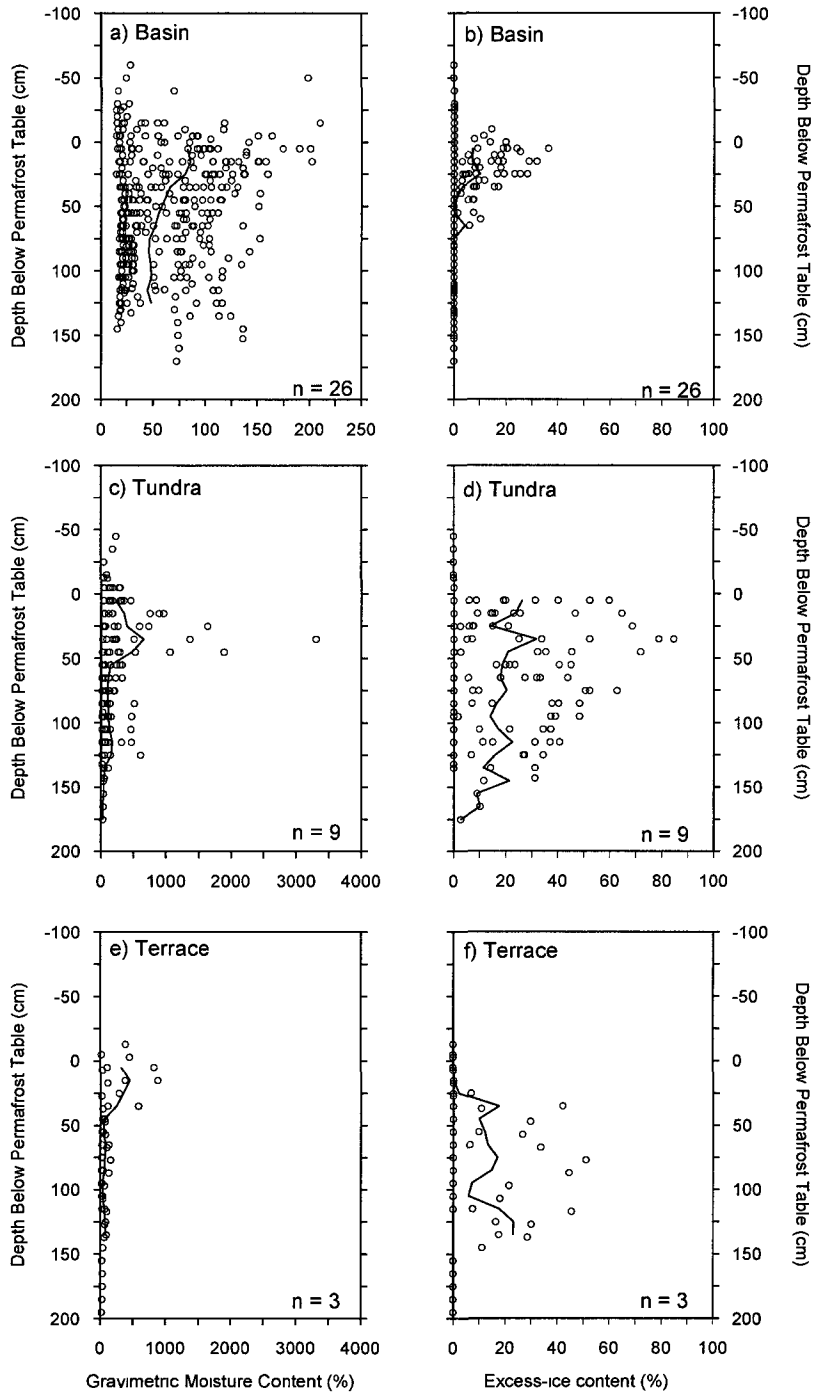


Figure 4.10 Gravimetric moisture contents for (a) 26 basin, (c) nine tundra and (e) three terrace sites and excess-ice contents for the same (b) basin, (d) tundra and (f) terrace sites. The data are plotted against the depth below the 2010 permafrost table. Circles represent the measured value of each 10 cm core segment. The solid black lines are the mean values of the circles between 0 and 100 cm below the permafrost table. The line does not extend beyond 100 cm because the number of samples at greater depths is limited.

Table 4-7 Excess-ice contents for the top 50 cm of permafrost, total ice accumulation in the upper metre of permafrost, and the accumulation rate over 32 years since drainage for all basin sites.

Basin Site	Excess ice in top 50 cm (%)	Ice accumulation in top 100 cm (cm)	Accumulation rate (mm y^{-1})
WW			
1	12.5	5.6	1.6
2	10.9	6.3	1.8
3	1.3	0.5	0.2
4	8.2	3.2	0.9
5	12.5	6.7	1.9
6	9.0	4.0	1.2
7	19.4	9.7	2.8
8	5.9	2.7	0.8
BC			
9	0.0	0.0	0.0
10	7.4	3.2	0.9
11	1.4	0.7	0.2
12	0.0	0.0	0.0
13	10.6	4.8	1.4
14	0.0	0.0	0.0
15	0.0	0.0	0.0
16	0.9	0.8	0.2
17	12.5	5.6	1.6
18	1.7	0.8	0.2
LM			
19	0.0	0.0	0.0
20	0.0	0.0	0.0
21	8.6	3.5	1.0
22	0.0	0.0	0.0
23	0.0	0.0	0.0
24	0.0	0.0	0.0
25	1.7	0.8	0.2
26	5.3	2.4	0.7

*Site 26 not included in basin unit classification due to thick peat accumulation at surface

active-layer was commonly saturated at these sites and water seeped into many of the bore holes during drilling. The soil at the top of permafrost at eight WW sites had mean silt content of 72% (range = 43.3-82.7%; n = 8) and mean organic matter content of 13% (range = 4.7-17.8%; n = 8), respectively. The mean excess-ice content of the upper 50 cm of permafrost was greatest at these sites (10%) (Figure 4.11b).

4.5.1.2 Basin centre

Ten drill sites were classified as BC (Table 4-7). The basin centre sites were generally much drier than WW sites. Soil at the top of permafrost was similar, with mean silt and organic matter contents at ten sites of 75% (range = 71.7-78.5%; n = 10) and 16% (range = 4.0-24.5%; n = 10), respectively. Mean excess-ice content of the upper 50 cm of permafrost at these sites was 4%, much lower than at wet willow sites (Figure 4.11d).

4.5.1.3 Lake margins/little vegetation

Seven drill sites were located at margins or in areas with little vegetation (Table 4-7). These sites were generally characterised by limited vegetation cover and deep, dry active layers. However, Site 24, immediately beside the lakeshore, had willows growing in excess of 3 m high. The soil at the top of permafrost was much sandier at seven LM sites than elsewhere, with mean silt and organic matter contents of only 41.3% (range = 16.3-77.0%; n = 7) and 4.0% (range = 2.4-9.0%; n = 7), respectively. Low soil organic matter contents resulted in substantially lower GMC at LM sites compared to WW and BC (Figure 4.11a,c,e). Mean excess-ice content of the upper 50 cm of permafrost was 1.7% (Figure 4.11f).

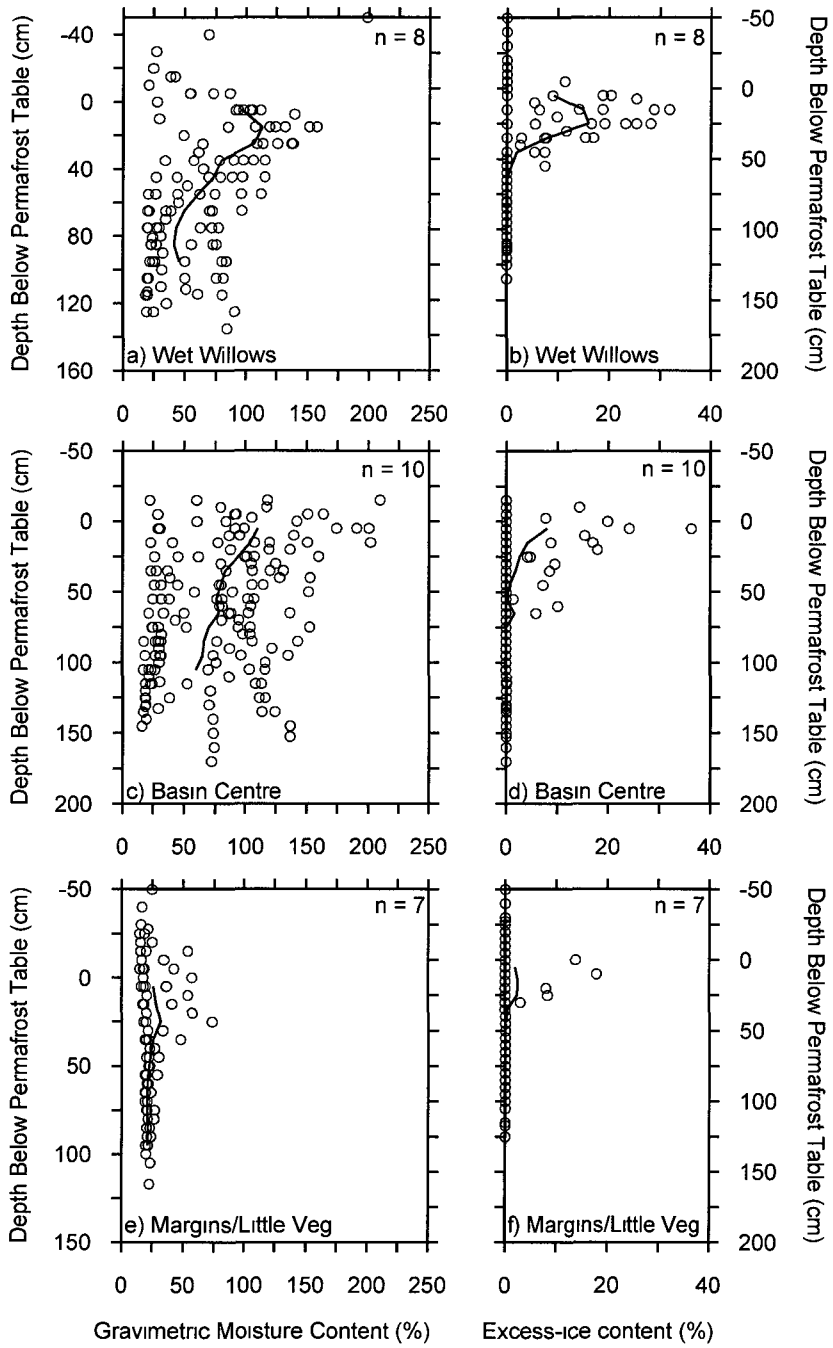


Figure 4.11 Gravimetric moisture contents for (a) eight wet willow sites, (c) ten basin centre sites and (e) seven lake margins/little vegetation sites, and excess-ice contents for the same (b) wet willow, (d) basin centre, and (f) lake margins/little vegetation sites. The data are plotted against the depth below the 2010 permafrost table. The circles represent the values for each 10 cm core segment, and the solid black line is the mean value plotted for depths between 0 and 100 cm below the permafrost table. The line does not extend beyond 100 cm because the number of samples at greater depths is limited.

4.5.2 Tundra

Near-surface ground ice was examined at nine sites in the tundra surrounding the Illisarvik lake basin. Excess-ice content of the upper 50 cm of permafrost was highly variable at tundra sites, between 4 and 50%. The mean excess-ice content of the top 50 cm of permafrost was 25%, much higher than at basin sites (Figure 4.10d). Excess-ice was also found much deeper below the permafrost table at tundra sites, to 1.5 metres below the 2010 active-layer depth (Site 28). This core was ice rich to its base, and so the ice enrichment may extend beyond 1.5 m below the permafrost table. The mean excess-ice content of the top 100 cm of permafrost at tundra sites was 20%, similar the top 50 cm. This similarity highlights how excess-ice exists at much greater depths in the tundra compared to the basin, where all enrichment is in the upper 60 cm of permafrost. The mean ice accumulation in the top 100 cm of permafrost at tundra sites was 17.8 cm, far greater than the mean at all basin sites (2.4 cm).

The maximum excess-ice content in the top 50 cm of permafrost occurred at Site 31, near the Illisarvik lakeshore at the base of a gentle slope. The top 40 cm of permafrost appeared as almost pure ice, and had mean excess-ice content over 60%. This may be evidence of the existence of large ice lenses at slope bases, which has been reported by Morse et al. (2009), or may be evidence of injection ice. Sites 32 and 28 had the lowest excess-ice contents in the top 50 cm of permafrost, and both were located at upland tundra areas. These results indicate that topographically controlled soil moisture supply may be important to the accumulation of ice at tundra sites.

4.5.3 Lake basin terraces

Three sites on former lake basin terraces were drilled to determine near-surface ground ice conditions. The mean excess-ice content of the top 50 cm for these sites was 6%. At all terrace sites, the majority of the ice enrichment occurred deeper in the core, and the mean excess-ice content for all terrace sites was 9% in the top 100 cm of permafrost (Figure 4.10f). Excess-ice was present at least up to 1.4 metres below the 2010 permafrost table, and extended to the bottom of the core at Site 38. The mean ice accumulation in the upper 1 m of permafrost was 9.9 cm, which is between the values obtained for the basin and tundra sites.

Absence of excess-ice at the very top of permafrost at terrace sites was likely due to high organic matter content. There was no excess ice in the top 20 cm of permafrost (Figure 4.10f), where organic matter content was very high. Excess-ice content was determined by melting the core segment and measuring the supernatant water volume, but the high porosity of organic matter accommodated the melted ice, leading to an excess-ice content of zero. This effect is well demonstrated at Site 37. The top of permafrost consisted of ice bonded peat, underlain mainly by silt or sandy silt. The ice-bonded peat of the upper 20 cm of permafrost had 800% moisture content, contained approximately 275 ml of water in the 20 cm core, and no excess ice. In contrast, from 50 - 70 cm below the permafrost table in silt soil, the moisture content was 90%, there was about 100 ml of water in the 20 cm core, and an excess ice content of 8%. Therefore, despite containing more than twice the amount of ice as the silty section of core, there was no excess ice in the top of permafrost due to the high porosity and water retention of peat.

To summarize, excess-ice contents in the basin were highly variable; the highest ice contents were in WW areas, while LM sites were the least enriched with ice. Much more excess ice has accumulated in the tundra surrounding Illisarvik, and high ice contents were observed at greater depths in the tundra than in basin sediments. On the lake terrace, excess-ice was absent from the top 20 cm of permafrost, likely due to high organic matter contents.

4.6 Gravimetric moisture content

The 1979 and 1980 core data from Michel (1982) are reported as gravimetric moisture content profiles. Comparison between the current moisture content profiles and those collected by Michel (1982) follow later in this chapter. This section provides a brief summary of gravimetric moisture content profiles from the basin, tundra and terrace.

4.6.1 Lake basin

Gravimetric moisture content was examined at the 26 lake basin sites (Figure 4.10a). Organic matter content accounted for most of the variability in moisture content at the top of permafrost ($r^2 = 0.77$, $p < 0.01$) due to its influence on the soil bulk density. Gravimetric moisture content was also associated with the excess-ice content at the top of permafrost ($r^2 = 0.37$, $p < 0.01$). The moisture content in the upper 50 cm of permafrost is visibly higher than below as a result of excess-ice accumulation (Figure 4.10a & b)

4.6.2 Tundra

Tundra moisture content was highly variable, especially at the top of permafrost where some values were an order of magnitude greater than those observed at basin sites

(Figure 4.10c). This is due to an organic-rich layer commonly found at the top of permafrost below earth hummocks. For example, the organic matter content 25 cm below the permafrost table was 82% at Site 29, with resulting moisture contents up to 3300% in the top 50 cm of permafrost.

4.6.3 Lake basin terraces

Gravimetric moisture contents observed in lake basin terraces were of similar magnitude to those seen at tundra sites (Figure 4.10e). The terraces are composed of thick layers of peat underlain by fine sands and silts. High moisture contents at Site 37 were because the peat layer extended into permafrost. The peat at Sites 36 and 38 was mostly in the active layer, resulting in lower moisture content at the top of permafrost at these two sites.

In summary, basin moisture contents were statistically related to organic matter content and excess-ice content. Soils in the basin were much more homogenous than in the tundra and terrace, which resulted in relatively limited variation in gravimetric moisture content. In the tundra and terrace, organic matter contents were highly variable and spatially heterogeneous, due to variations in organic matter content beneath tundra earth hummocks, and the peat layer on the lake terrace.

4.7 Relations between basin ice accumulation and potential controlling factors

Data collection and site characterization was designed in order to examine relations between near-surface ground ice accumulation and several potential controlling factors. Sites were classified as either wet or dry to examine soil moisture status as a

control on ice accumulation. Soil texture was determined for samples taken from the upper permafrost to observe relations between soil type and the amount of near-surface ground ice. Soil organic matter content at the top of permafrost was determined to examine its association with ice accumulation in the basin. Vegetation height was measured to assess whether snow trapping and subsequent warming of the permafrost affected ground ice accumulation.

4.7.1 Soil moisture status

Of the 26 primary drill sites in the basin, thirteen were in wet areas, and thirteen in dry areas. The majority of wet sites were located around North Pond in soggy ground with willows. The others were located in wet areas near the basin centre and at the southern end of the basin adjacent to wet depressions. Dryer sites were predominately in grassy areas near the basin centre or around the southern lake margins. Site locations are presented in Figure 4.9. The excess-ice content in wet areas was significantly higher than in dryer areas [Mann-Whitney $U(13) = 39$, $p < 0.01$]. The mean excess-ice content of the upper 50 cm of permafrost was 7% at wet sites and 3% at dry sites. Eleven of the wet sites contained excess-ice in the top 50 cm of permafrost, whereas only six of the dryer sites were enriched with ice (Figure 4.9).

4.7.2 Soil texture

The resulting particle size distributions obtained from laser diffraction were divided into three texture classes (2 mm > sand > 53 μm > silt > 2 μm > clay > 0.36 μm) (USDA classification). Silt was the dominant soil texture class at the top of permafrost at most basin sites. The median silt content at 26 basin sites was 74%, while the median clay and sand contents were 18 and 7%, respectively. Sand was the dominant soil texture

class at only five sites in the basin. These five sites were the closest sites to the basin margins, all within 30 m of the former lakeshore.

Soil texture had a noticeable, but not statistically significant influence on ice enrichment at the top of permafrost. The excess ice-content of the top 50 cm generally increased with increasing silt content (Figure 4.12), but the relation between silt content and excess-ice was not statistically significant at the 95% confidence level ($r^2 = 0.09$; $p = 0.14$). The large variation in excess-ice contents at sites with high silt contents indicates that other factors are important to ice formation.

Excess-ice content at the top of permafrost decreased with sand content, but the relation was also not statistically significant at the 95% confidence level ($r^2 = 0.08$; $p = 0.16$) (Figure 4.12). The range of clay contents in the lake basin was limited compared to silt and sand, however excess-ice content of the top 50 cm of permafrost generally increased with increasing clay content (Figure 4.12). The observed relation between clay content and excess-ice at the top of permafrost was not significant at the 95% confidence interval ($r^2 = 0.03$; $p = 0.36$).

The relations between soil texture and gravimetric moisture content were also examined. The gravimetric moisture contents for each 10 cm core segment analyzed for soil texture are plotted against soil texture in Figure 4.13. There was a clear increase in gravimetric moisture content with increasing silt contents, and the relation was statistically significant at the 95% confidence level ($r^2 = 0.32$; $p < 0.01$). The relation between sand content and gravimetric moisture content was also significant ($r^2 = 0.31$; $p < 0.01$), with moisture content decreasing as the sand fraction increased. Gravimetric

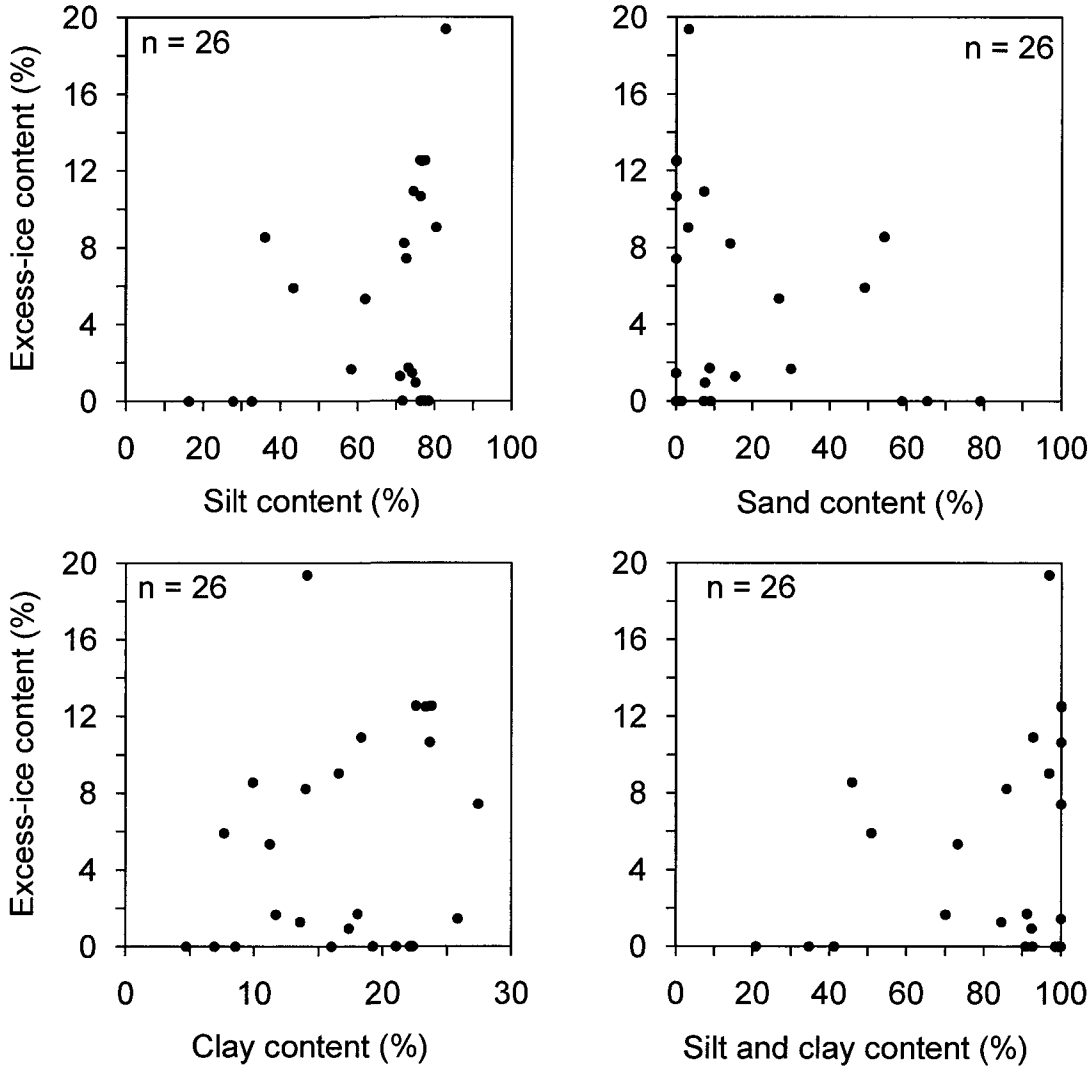


Figure 4.12 Relations between mean excess-ice content of the top 50 cm of permafrost at each basin drill site and silt ($r^2 = 0.09$; $p = 0.14$), sand ($r^2 = 0.08$; $p = 0.16$) and clay ($r^2 = 0.03$; $p = 0.36$) contents from one sample of the upper permafrost. Note the different scale of the clay content axis.

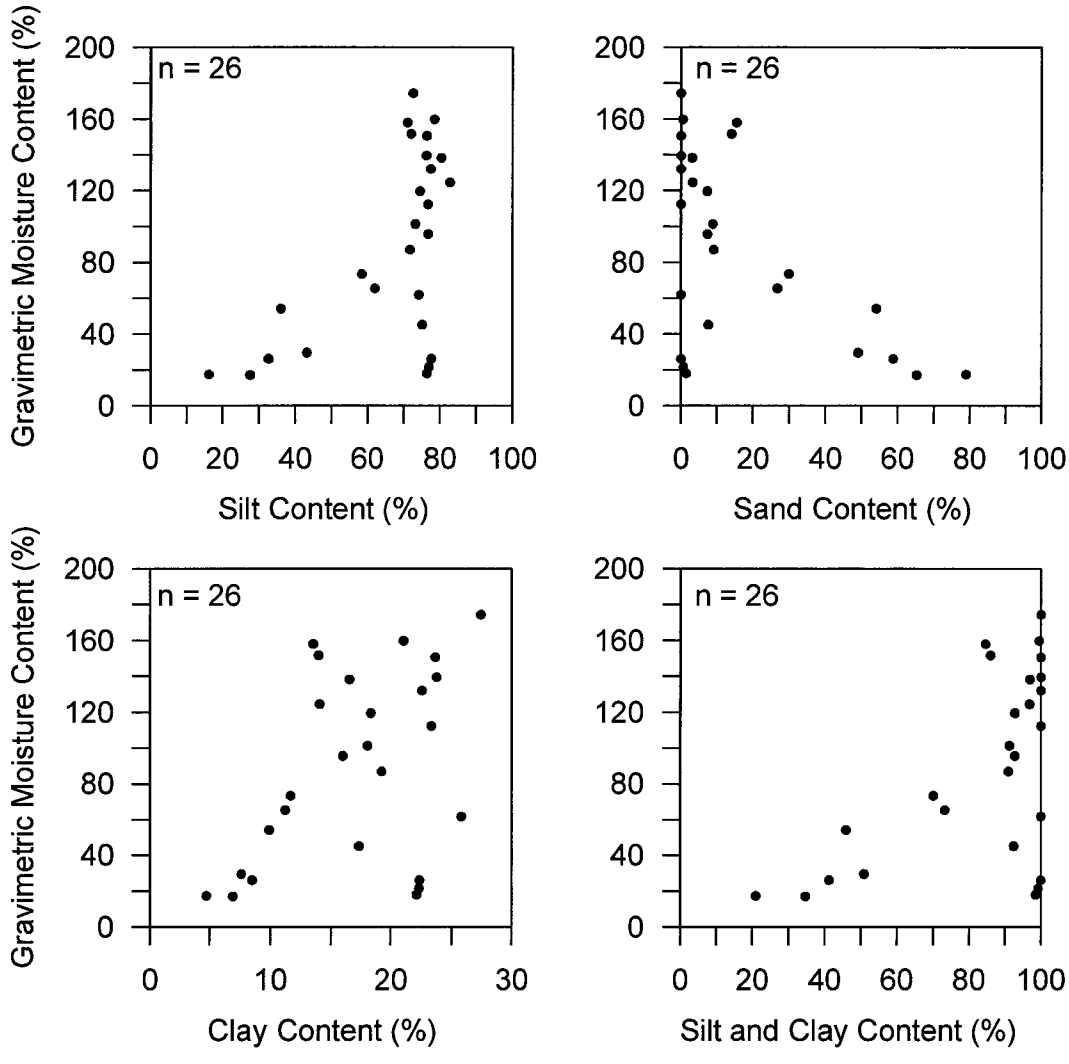


Figure 4.13 Relations between gravimetric moisture content of one 10 cm section from the upper 30 cm of permafrost and silt ($r^2 = 0.32$; $p < 0.01$), sand ($r^2 = 0.31$; $p < 0.01$) and clay ($r^2 = 0.19$; $p = 0.02$) content from the same core section at each basin site.

moisture content also increased with higher clay fractions. The association was not as clear as for sand and silt, but was also significant at the 95% confidence level ($r^2 = 0.19$; $p = 0.02$).

Gravimetric moisture content was statistically related to soil texture because soil texture influences bulk density. Bulk density increases with sand content. In addition, the sandiest sites had low organic matter contents and lacked excess-ice accumulation, which resulted in low moisture contents due to the high bulk densities and absence of excess ice. In contrast, silt-rich sites had more organic matter, lower bulk densities and more ice accumulation, resulting in higher gravimetric moisture contents. Excess-ice was not statistically related to soil texture likely because of the large variation in ice content observed at silt-rich sites.

4.7.3 Soil organic matter content

The soil organic matter content from one segment in the top 30 cm of permafrost was determined for each of the 26 basin sites. Organic matter content ranged between 2 and 25%, and the median was 12%. The lowest organic matter contents occurred at sites near the true south lake margins, in sandier soils, while sites with the highest organic matter contents were located near the deeper centre of the lake basin in areas dominated by silt (Figure 4.14). The relationship between organic matter content and distance from the shore is significant at the 95% confidence level ($r^2 = 0.44$; $p < 0.01$) (Figure 4.15). This may be attributable to development of organic sediments in deeper areas of the basin near the lake centre where ice does not reach the lake bottom in winter. Vegetation in shallower areas may be disturbed by freezing of sediment when ice reaches the lake bottom (e.g. Alasaarela et al., 1989).

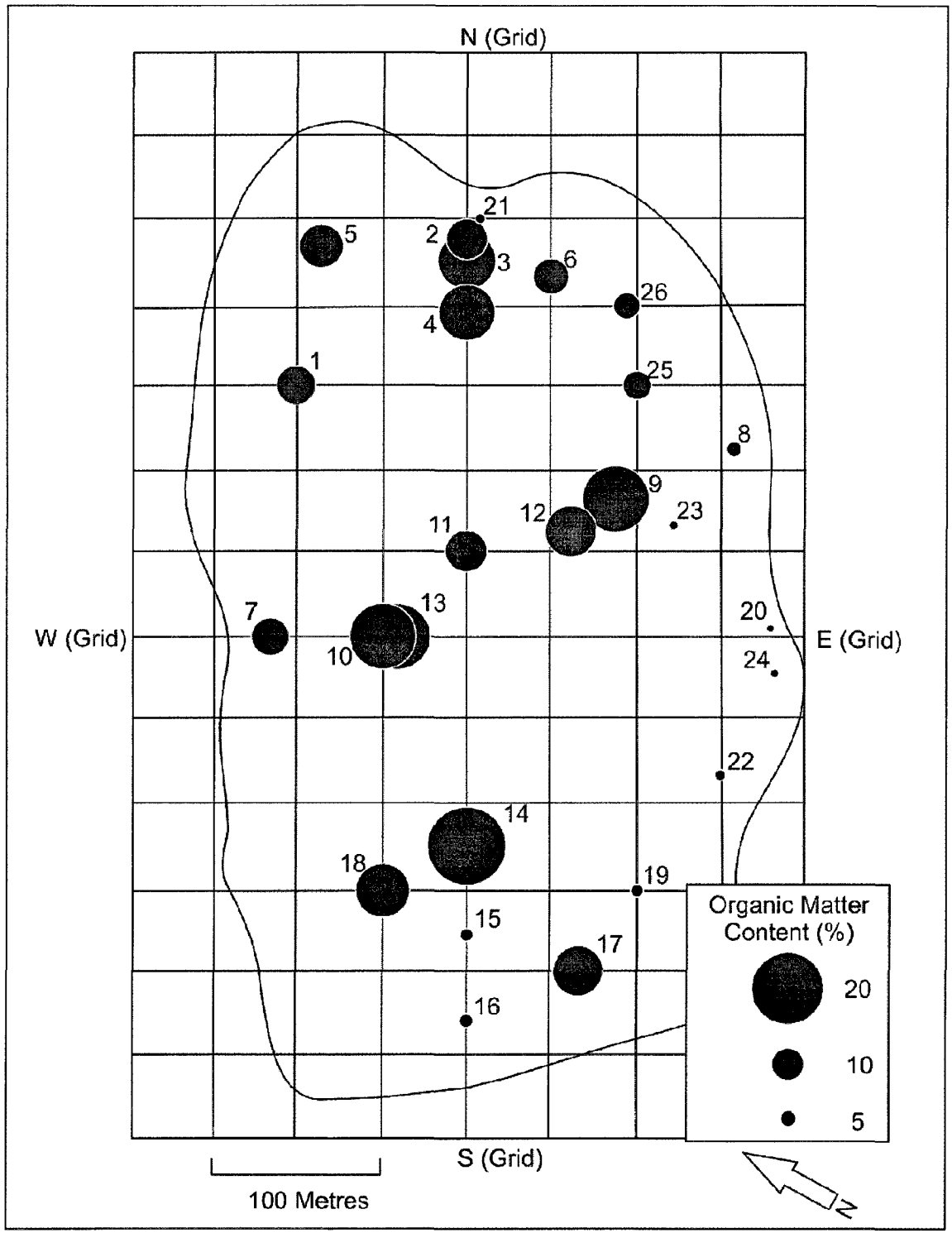


Figure 4.14 Proportional symbol map of organic matter content at the top of permafrost at basin sites. The radius of the circle is proportional to the organic matter content.

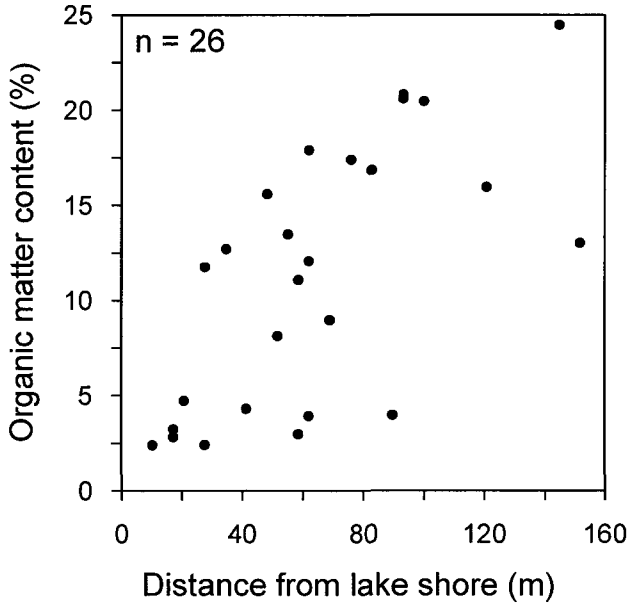


Figure 4.15 Relation between organic matter content at the top of permafrost and the distance of the drill hole from the lake shore ($r^2 = 0.44$; $p < 0.01$).

Soil organic matter was examined as a possible control on ice accumulation. There is no significant relation between excess-ice content of the top 50 cm of permafrost and soil organic matter at the 95% confidence level ($r^2 = 0.05$; $p = 0.251$) (Figure 4.16).

4.7.4 Vegetation height (temperature)

Vegetation growth and resulting snow entrapment lead to higher winter permafrost temperatures, facilitating migration of unfrozen water upwards out of near-0°C permafrost in winter (Cheng, 1983). On this basis, less ice at the top of permafrost would be expected at sites with higher vegetation. Vegetation heights were measured at each site to examine the effect of snow entrapment on ice enrichment. There was no observed relationship between vegetation height and near-surface ground ice accumulation ($r^2 = 0.02$, $p = 0.47$) (Figure 4.17). This suggests either that the ground temperature increase caused by snow entrapment has no effect on ice accumulation, or that the effect is masked by more influential factors. The small number of seasonal cycles since drainage may also restrict the impact of this consideration. Vegetation height was related to soil moisture status; wet sites had statistically higher vegetation than dry sites (Figure 4.18) [Mann-Whitney $U(26) = 42$, $p = 0.029$].

4.7.5 Summary

Soil moisture status was significantly associated with excess-ice accumulation. Soil texture also had a noticeable influence on ice enrichment, but the relation was not statistically significant, likely due to variation in ice content at sites with high silt contents. Neither organic matter nor vegetation height was associated with excess-ice accumulation in the basin. Organic matter content in sediments increased with the

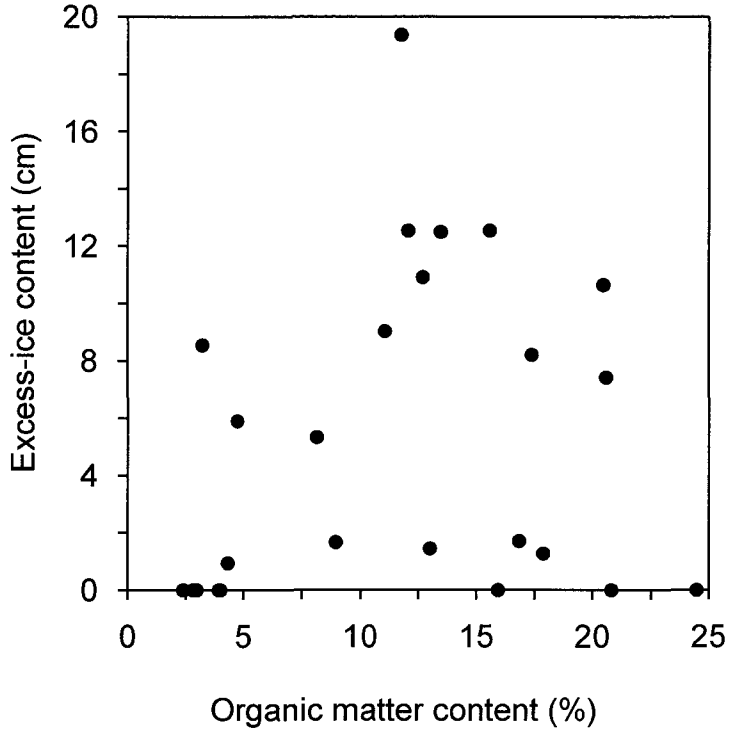


Figure 4.16 Relation between organic matter and excess-ice content of the top 50 cm of permafrost at 26 basin sites ($r^2 = 0.05$; $p = 0.25$).

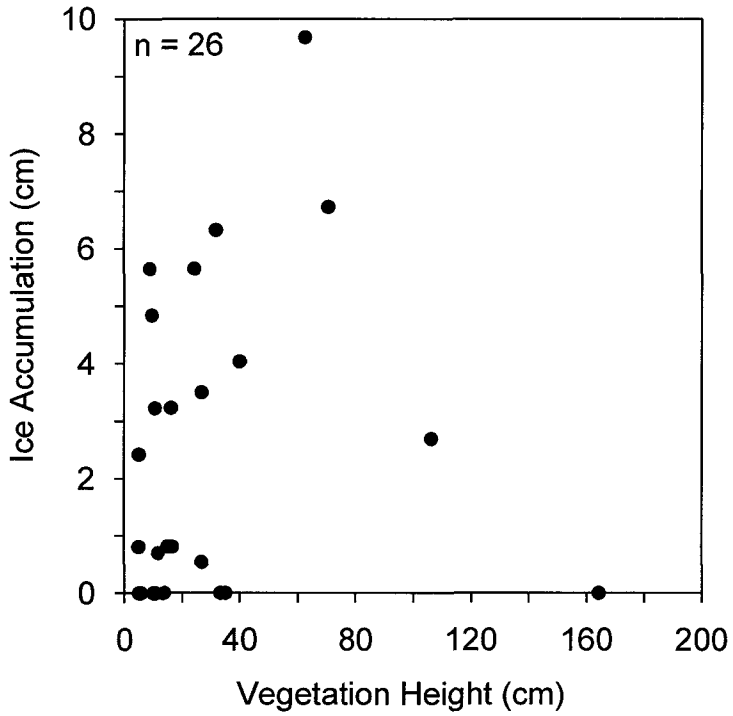


Figure 4.17 Ice accumulation and mean maximum vegetation height at 26 basin sites ($r^2 = 0.02$; $p = 0.47$).

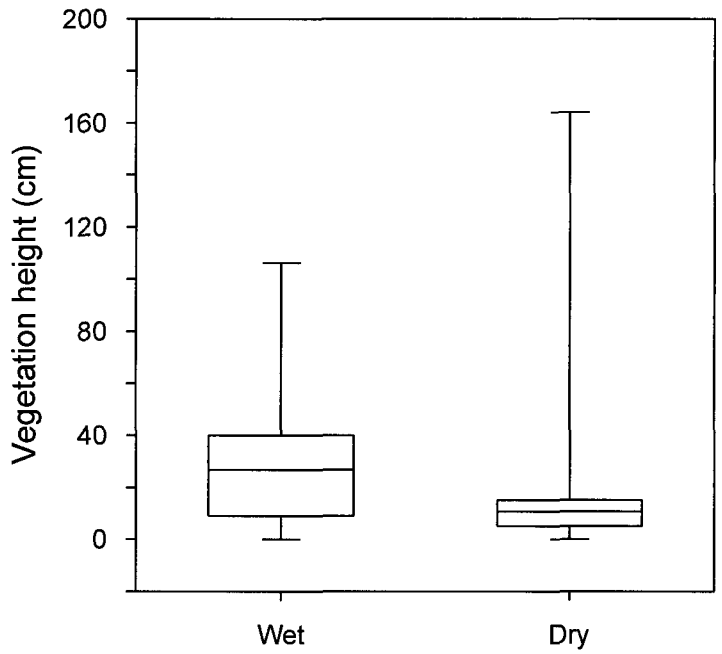


Figure 4.18 Vegetation heights at wet and dry basin sites (n = 13 for each class). The horizontal line inside the box represents the median vegetation height, the top and bottom of the box are the upper and lower quartiles, and the whiskers are the maximum and minimum extent of vegetation heights at the sites.

distance of the drill site from the former lake shore, likely because winter ice cover that reached the bottom in shallower areas disturbed vegetation growth.

4.8 Evidence of moisture migration to the top of permafrost

Sites drilled by Michel (1982) were revisited in this study to assess change in moisture content over 30 years. In the basin, comparisons at only two sites were possible. In lakeshore terraces and the tundra, comparisons were only possible at one site. The number of comparisons was limited because most sites drilled by Michel (1982) were not analyzed for moisture content. Of the sites where comparison was possible, Site 12 was not enriched with excess ice, and Site 18 contained little excess ice at the top of permafrost. These two sites are located in areas where ice enrichment is generally limited, and are not representative of the entire basin.

At Site 18 in the lake basin, moisture content changed a little with depth below the 2010 permafrost table (Figure 4.19a). Between 50 and 100 cm, moisture content decreased significantly in the past 30 years, and increased between about 110 and 150 cm. This may be evidence of a net downward water migration into permafrost since 1979. Conversely, the differences may in part be a result of differing organic matter contents between the drill sites. The organic matter content at the top of permafrost was 17%, so if the organic matter content at the location drilled in 1979 was lower, this may account for the difference. Excess ice was only present between 80 and 90 cm; no measurable excess-ice enrichment occurred between 110 and 150 cm where the moisture contents were greater. Below 150 cm, moisture contents did not change noticeably between 1979 and 2010. At Site 12 in the basin, the moisture contents of the top metre of permafrost

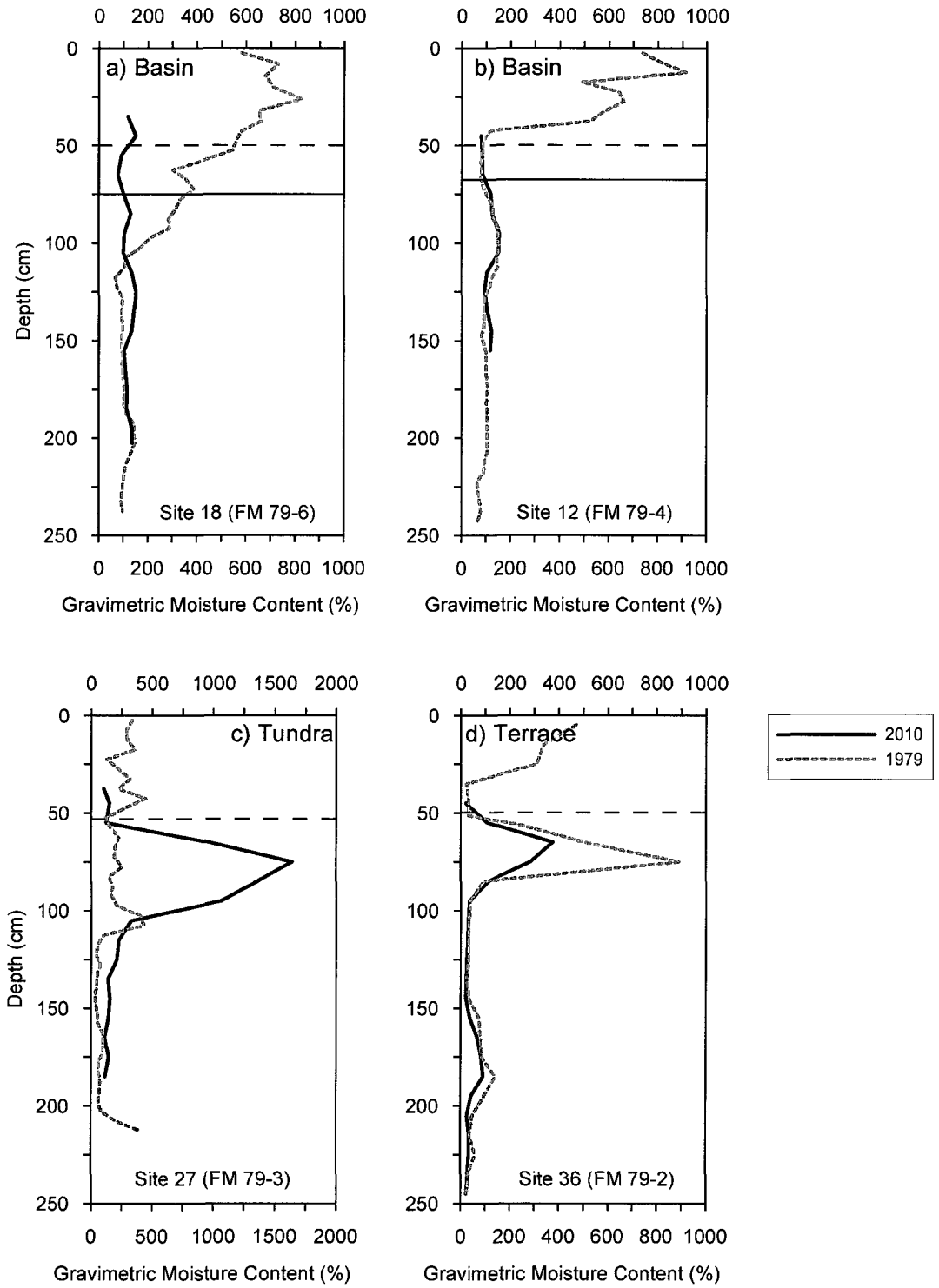


Figure 4.19 Gravimetric moisture content curves for 1979 and 2010 drill sites. The dashed horizontal line represents the 2010 permafrost table, and the solid horizontal line represents the 1979 permafrost table from nearby benchmarks in the basin.

remained essentially unchanged, suggesting that there was little increase in ice content during permafrost aggradation (Figure 4.19b).

In comparison with data collected in 1979, the moisture content at Site 27 in the tundra was substantially different in 2010 (Figure 4.19c). There was a large peak in moisture content in 2010 just below the contemporary permafrost table. This is not likely indicative of moisture migration to the top of permafrost, but rather of differing organic matter conditions. The hummocky tundra terrain is morphologically heterogeneous, and though the drill sites were close together, variation in organic matter content at the top of permafrost likely resulted in the observed differences. This variation in organic matter content over small spatial scales limits the value of repeat drilling in tundra for temporal comparisons of moisture and excess-ice content.

Similar to the tundra site, the moisture content at lake terrace site 36 was significantly different at the top of permafrost (Figure 4.19d). The moisture contents measured in 1979 were much higher than those of 2010, to approximately 30 cm below the permafrost table. Drill logs and loss-on-ignition indicate that the core was highly organic (45%) at the top of permafrost. Small differences in ice content between the drill sites can result in large differences in moisture content due to the low soil bulk density. At depths greater than 80 cm, the moisture contents at Site 36 remained essentially unchanged between measurement periods.

The results of these comparisons are inconclusive and do not provide evidence of moisture migration to the top of permafrost. This may be largely due to the lack of data

from 1979, and the limited number of seasonal cycles allowing moisture to accumulate at the top of permafrost.

5. DISCUSSION

5.1 Introduction

In Chapter 4, the varying responses of basin, tundra and terrace active layers to seasonal air temperatures and winter snow depths were examined. Ground ice accumulation below the active layer was highly variable in the basin, with the majority found in wet willowed areas. The upper permafrost had much more ice in the tundra than in the basin and terrace. This chapter discusses near-surface ground-ice formation and active-layer development in further detail. Specifically, this chapter will: (1) discuss the factors associated with variation in active-layer depths in the basin, tundra and terrace; (2) assess the relative importance of controlling factors on ice accumulation; (3) discuss rates of near-surface ground ice accumulation; (4) assess the mechanisms of near-surface ground ice formation in the basin, tundra and terrace; and (5) discuss the transient layer conceptual model.

5.2 Active-layer development

Active layers vary with air temperature, vegetation, snow cover, substrate type, slope, aspect, and soil-water content (Mackay, 1982; French, 2007). Active-layer variations and their relations with air temperatures and snow depths over the past 30 years were examined for the lake basin, tundra and terrace. In the flat lake basin, slope and aspect are insignificant contributors to variation in active-layer thicknesses. Basin active layers in the basin centre (BC) and near the lake margins and areas with little vegetation (LM) were significantly correlated with summer air temperatures during the

1980s when vegetation was absent or sparse in these areas. The correlations became weaker in subsequent decades, concurrent with vegetation establishment (Mackay and Burn, 2002b). Active layers in wet willow areas (WW) were not significantly associated with summer temperatures in the first two decades, likely because of the wet ground and subsequent early vegetation establishment in these areas altered the summer surface energy balance in the years immediately following drainage.

The active layers in WW did not increase during the 1990s as they did in dryer, less-vegetated BC areas. This may have resulted from the vegetation, soil moisture and ground-ice characteristics in the densely willowed areas. Insulation and shading by the vegetation may have reduced the amount of summer energy available to the active layer. In addition, the willow sites were located in very wet ground, where abundant ice lenses may have formed in the active layer during freezeback (e.g. Mackay, 1981). The latent heat required to melt ice lenses (334 kJ/kg) is substantial and may have limited thaw progression in the active layer (Romanovsky and Osterkamp, 1997; Zhang and Stamnes, 1998), reducing the statistical associations with seasonal temperatures. The higher amount of near-surface ground ice in the permafrost at WW sites also likely protected subjacent permafrost from thaw (Shur et al., 2005).

Significant thinning of the active layers observed in WW and BC between the 1990s and 2000s suggests that summer insulation from increased vegetation cover was more important for changing thaw depths than permafrost warming from increased snow entrapment in the basin, but cooler spring temperatures in the 2000s compared to the 1990s may have also contributed to the thinning.

Snow depths were significantly correlated with basin active-layer thicknesses at the BC sites. The associations in the other basin classes were not significant, but this may be in part due to the limited number of observations (n = 23). Multiple regressions using the stepwise method indicated that over the entire period of record, between 31% (LM) and 61% (BC) of the variability in basin active-layer depths was accounted for by seasonal temperatures and snow depths (Table 4-4). Summer temperature was the only independent variable that significantly influenced the models for LM sites. Summer temperatures and snow depths both had a significant influence on the models in WW and BC units.

The low r^2 values from multiple regressions for LM and WW sites suggest that soil moisture and vegetation effects may be important to active-layer variation in the basin. The importance of these variables to active-layer thicknesses has been observed in several studies (e.g. Price, 1971; Mackay, 1995; Nelson et al., 1997; Zhang and Stamnes, 1998). Soil moisture and vegetation conditions were unique in WW and LM, which may have resulted in the lower r^2 values. The low values may have also been a result of variation in site conditions within each basin unit. The aggregation of active-layer thicknesses and snow depths into the basin units may have reduced the associations between these variables.

Active layers at sites in BC had a stronger association with summer temperatures and snow depths, likely because the sites were generally dryer, less vegetated, and had limited near-surface ground ice at the top of permafrost. The reduced effects of soil moisture, insulation and shading from vegetation in the summer, and latent heat

requirements of the active-layer and upper permafrost likely explain the greater association between active-layer depths and air temperatures and snow depths.

On the tundra hillslope, active-layer thicknesses have not changed significantly since data collection began in 1983. However, active layers have been slowly increasing in thickness on the tundra hilltop. The hillslope sites were generally much wetter than hilltop sites, and had more excess-ice at the top of permafrost. The mean excess-ice content at three sites (sites 27, 30, 31) on tundra slopes was 35%, whereas the mean was 15% at four tundra hilltop sites (sites 28, 32, 34, 35). The high ice contents at the top of permafrost may currently limit active-layer thickness increases on the tundra hillslope due to the latent heat required for thaw.

Multiple regressions of seasonal air temperatures and snow depths and active-layer depths in the tundra provided results comparable to those observed in WW, BC, and LM areas of the lake basin. Summer temperatures accounted for 55% of the active-layer variability at hillslope sites and 36% at hilltop sites. The summer temperature was the only independent variable that significantly influenced the regression model on the hillslope and hilltop.

The active-layer depths at benchmark 41 on the lake terrace have not changed significantly since 1979. The stable permafrost table, as mentioned in Chapter 4, is likely a result of the unique surface conditions on the terrace. The thick peat layer may contain substantial ice due to its high porosity, as was observed at Site 37 (Section 4.5.3), and represent a major latent heat requirement in summer during active-layer thaw. In addition, peat at the surface may dry in the summer and effectively insulate the ground

below due to its low thermal conductivity. Highlighting this effect, Fitzgibbon (1981) measured thermal conductivities in dry peat about 30 times less than in wet peat in central Saskatchewan. The multiple regression for benchmark 41 highlights the unique surface and soil characteristics of the basin terrace. The snow depths and seasonal temperatures accounted for only 9% of active-layer variation, suggesting that the active layer is insensitive to air temperatures and snow depths.

Another important result from benchmark 41 on the lake terrace was the unique snow depth record. While snow depths have increased statistically significantly since the 1980s in both the lake basin and tundra, there was no significant trend at Benchmark 41, where there is no shrub vegetation. This suggests that increases in vegetation height may be resulting in greater snow entrapment in the basin, which was reported previously by Mackay and Burn (2002a), and in the tundra, which has not been reported from the study site.

5.3 Controlling factors of ice accumulation

5.3.1 Soil moisture status

The assessment of controlling factors of near-surface ice accumulation in the Illisarvik lake basin yielded several conclusions. First, persistent soil moisture status was likely the most important controlling factor of near-surface ground ice formation in the lake basin. This was evident by the contrasting ice contents of drill sites in wet and dry areas (Figure 4.9). Six of thirteen sites in dry areas were enriched with excess-ice, compared to eleven of thirteen sites in wet areas. There was much more near-surface

ground ice in WW areas compared to the other basin classification units (Figure 4.11). The amount of ice present in WW compared to BC perhaps best demonstrates the importance of soil moisture to near-surface ground ice formation. The soils in both units were frost-susceptible and had similar mean silt contents near 75%. Active-layers in all three units have thinned since the 1990s, favourable to aggradational ice formation. Despite these similarities, mean excess-ice content in the top 50 cm of permafrost in WW (10%) was more than twice that in BC (4%). The importance of water availability to the formation of near-surface ground ice has been observed in previous studies. Kokelj and Burn (2005) observed greatest ice accumulation where permafrost had aggraded in a saturated environment, while Burn (1988) measured rapid ice formation in saturated soil tubes.

5.3.2 Soil texture

Soil texture was also a control of ice accumulation at the top of permafrost. The excess-ice content was generally greater at sites with higher silt contents, and was less at sites with higher sand contents. No sites with greater than 55% sand content contained excess ice. Soils at sites 8 and 21 had relatively high sand contents of 49 and 54%, respectively, but both were enriched with excess-ice. These sites had near 40% silt content and were therefore still frost-susceptible despite the relatively high sand contents.

Six of the 26 basin drill sites had high (> 40%) silt contents at the top of permafrost but were not enriched with excess ice. Sites 9, 15, 19, and 23 were all located in dryer areas, so moisture may have been a limiting factor to ice accumulation. In addition, Site 23 was located in a dry, bare area where the active-layer depth was considerable (120 cm), so the lack of aggradation of the permafrost table may have also

limited excess-ice formation. Sites 12 and 14 had frost susceptible soils, were located in wetter areas, and have had an aggrading permafrost table, but had no excess-ice accumulation. The lack of ice at these two sites is difficult to explain, as all conditions seemed favourable to aggradational ice formation.

5.3.3 Active-layer history

Results from this study have shown that basin active layers have generally been thinning since maximum thaw depths were observed in 1989 or 1998. Measured active layers in WW were reduced by 7 cm between the 2000s and 1990s, and in BC, mean active-layer depths in the 2000s were 9 cm thinner than in the 1990s. The exception to this thinning has been at sites near the lake margin and in areas with little vegetation, where depths remained similar in the past three decades (Figure 4.4).

In the tundra, the active layer has likely thinned slowly since the early-Holocene climatic optimum around 8000 years ago, when the active layer was probably 2.5 times its current thickness (Burn, 1997). Results from this study have shown that in the past few decades, tundra active-layer depths at hilltop sites have been increasing slowly. Mackay and Burn (2002a) similarly reported that tundra active-layer thicknesses between 1983 and 1999 had increased in response to higher summer temperatures.

Results from WW and BC suggest that active-layer history has been of little importance to ice formation in the lake basin. It is expected that more excess-ice would be present where the permafrost table aggradation is greatest. Active layers in BC have thinned on average by 9 cm since the 1990s, but the excess-ice accumulation was much lower than in WW, where active layers have thinned by 7 cm. These results suggest that

the influence of permafrost table aggradation is small in the basin, because moisture and soil characteristics have dominated as controls of aggradational ice formation. Aggradation of the permafrost table in the basin has also taken place over a short period, limiting the process of repeated-segregation due to the small number of seasonal cycles (Cheng, 1983).

The importance of active-layer history over longer periods is evident when comparing ice conditions in the basin and tundra. The overall slow and substantial rise of the permafrost table in the past 8000 years is likely the cause of high excess-ice contents and substantial ice enrichment at depth. The rise of the permafrost table would have enabled trapping of ice lenses at the base of the active layer to form aggradational ice, while the slow thinning would have also allowed substantial repeated-segregation ice to form from thousands of seasonal cycles transporting moisture to the top of permafrost.

5.3.4 Organic matter content

There was no relation between organic matter content at the top of permafrost and excess-ice content at basin sites. However, by examining organic matter content as a control on ice formation, another interesting relation was observed. There was a strong and significant relation between organic matter content and the distance of the drill site from the lake shore. This may be because deeper areas of the lake near the basin centre were not frozen to the bottom during winter, enabling aquatic vegetation to persist undisturbed. If this is the case, then the relation between organic matter content and distance from shore should be observable in other drained lake basins where the lakes were deeper than winter ice thicknesses.

5.3.5 Vegetation height (temperature)

Vegetation height was not related to ice accumulation at basin drill sites. Though many of the most ice-rich sites were in areas with abundant willow growth, the tallest vegetation was found at Site 24 at the lake margin, where sand was the dominant soil type, the active layer was very dry, and no excess-ice was measured. There were also many sites with shorter vegetation that had substantial ice accumulation. As a result, vegetation height was not indicative of ice conditions, and any effects on seasonal moisture migration and repeated-segregation ice formation from differences in ground temperature due to snow entrapment were likely masked by more important factors, such as moisture availability and soil texture. Again, the small number of seasonal cycles since drainage has probably limited repeated-segregation ice formation (Cheng, 1983), making the effect of permafrost temperature on ice accumulation negligible.

5.4 Rates of near-surface ground ice accumulation

At ice-rich sites in the basin, rates of near-surface ground ice accumulation were between 0.2 and 3.0 mm yr^{-1} over 32 years since drainage. The highest rates were in the wet, willowed areas surrounding North Pond. Total ice accumulation in the top 100 cm of permafrost was between 6 and 10 cm at these sites. Mackay and Burn (2002a) reported around 5 mm yr^{-1} of aggradational ice growth over 20 years at similar sites in the willows near North Pond. The total ice accumulation, inferred from measurements of benchmark protrusions, was between 8 and 12 cm (Mackay and Burn, 2002a), suggesting current ice contents around North Pond are similar in magnitude. The reported rates of accumulation

in willow areas from data collected in 2010 are likely lower because the ice accumulation is integrated over a longer time period.

The results from this thesis provide further evidence that rates of near-surface ground ice accumulation decrease over time (Kokelj and Burn, 2003). Figure 5.1 shows the relationship between accumulation rates and the time period of accumulation for all known studies. The maximum accumulation rates reported in each study are plotted, so that rates for the most favourable ice formation conditions are represented. It appears that the ice accumulation rate declines exponentially with time, though observations are limited in number. Further ground ice studies examining accumulation rates over different time periods may continue to strengthen this hypothesis.

5.4.1 Implications of reporting accumulation rates

Reporting the formation of near-surface ground ice as a rate may be of limited use in comparing ice accumulation between studies. Comparisons between different studies may only be useful if the periods of near-surface ground ice formation are equal, due to the seemingly exponential decrease in rates over longer periods. In addition, when active-layer depths over the entire period of near-surface ground ice formation are available, as in this study, reporting ice accumulation as a rate over the total ice formation period may not be desirable. For example, even though near-surface ground ice has been accumulating in the lake basin for 32 years since drainage, maximum active-layer depths at most sites were observed in 1989 or 1998. Therefore, much of the excess ice formed at the top of permafrost before these years of maximum thaw would have melted and reformed as the permafrost table subsequently rose in following years.

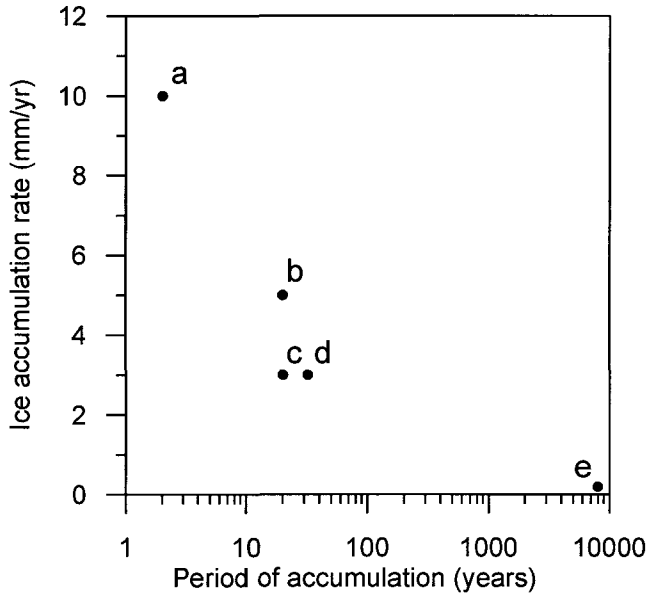


Figure 5.1 Near-surface ground ice accumulation rates for (a) soil tube experiments (Burn, 1988), (b) the Illisarvik drained lake basin (Mackay and Burn, 2002a), (c) ice formed above a 1981 thaw unconformity (Kokelj and Burn, 2003), (d) the Illisarvik drained lake basin (this study), and (e) ice formed above an 8000 year-old thaw unconformity (Burn, 1988).

In such a case, reporting the rate over 32 years would be an underestimation of the speed of ice accumulation.

5.5 Aggradational ice or repeated-segregation ice?

5.5.1 Evidence of repeated-segregation ice formation

Results from this study provided little or no evidence of moisture migration to the top of permafrost and subsequent accumulation of repeated-segregation ice in the lake basin since drainage. The 2010 gravimetric moisture content profiles at two lake basin sites did not indicate that moisture enrichment has occurred in upper permafrost since the sites were drilled by Michel (1982) (Figure 4.19). At Site 18, between 50 and 100 cm depth, which is currently the upper 50 cm of permafrost, the moisture content has declined since 1979. Moisture at these depths after initial freeze back of the lake basin surface may have been lost to the active-layer in following years. Site 12 shows no evidence of moisture migration to the top of permafrost over 32 years since drainage, and contained no excess-ice. However, it should be noted that these sites were located in areas with generally little near-surface ground ice, and therefore are not representative of the entire basin.

The lack of evidence for moisture migration to the top of permafrost and subsequent ice enrichment is likely due to the short time period since drainage. Because the migration of liquid water through frozen soil occurs very slowly, more seasonal cycles may be necessary to obtain measurable results. In addition, without accurate knowledge of the soil bulk density for each core segment in drill cores from both studies,

it is impossible to know whether an increase in moisture content is due to moisture migration to the top of permafrost or from differing organic matter contents or soil types at the drilling location.

Repeated-segregation ice has probably contributed significantly to near-surface ground ice in the tundra due to the slow aggradation of the permafrost table over the past several thousands of years. The higher excess-ice contents in the tundra compared to basin sites may be in part due to seasonal migration of moisture to the top of permafrost. The maximum excess-ice contents in basin core sections were around 40%, while in the tundra, excess-ice contents over 70% were observed. The average excess-ice content of the top 50 cm at tundra sites was 4.1 times greater than at basin sites. It seems that time has influenced near-surface ground ice accumulation, and repeated-segregation ice may be in part responsible for the drastic differences observed between tundra and basin sites.

5.5.2 Evidence of aggradational ice formation

The apparent dominance of active-layer soil moisture status on ice formation, the lack of evidence to suggest repeated-segregation ice development, and the limited number of seasonal cycles allowing moisture migration to the top of permafrost suggest that aggradational ice has been the principal mechanism of near-surface ground ice accumulation in the basin. Since aggradational ice development requires the trapping of ice lenses at the base of the active-layer, moisture conditions at the base of the active layer during freezeback must be favourable to ice lens growth. Very wet conditions were observed at the base of the active layer at some sites, particularly in WW areas. Commonly, water trickled from the base of the active layer into the boreholes, indicative of saturation at the active-layer base. With these moisture conditions and the rise of the

permafrost table between the 1990s and 2000s, the vast majority of ice in the basin is probably aggradational in origin.

Aggradational ice has likely been forming in the tundra since the early-Holocene climatic optimum around 8000 years ago (Burn, 1997). As the active-layer thinned progressively, ice lenses formed at the base of the active layer would have been trapped and incorporated into permafrost during upward freezeback. The presence of excess ice, measured up to 1.5 m below the permafrost table in this study, is indicative of active-layer thinning and aggradational ice formation. It is not possible to determine the relative amounts of aggradational and repeated-segregation ice in the tundra with the data collected for this thesis, but the magnitude of the active-layer thinning and the long time period over which the thinning occurred together suggest that both mechanisms of formation may be well represented.

5.6 The transition zone conceptual model

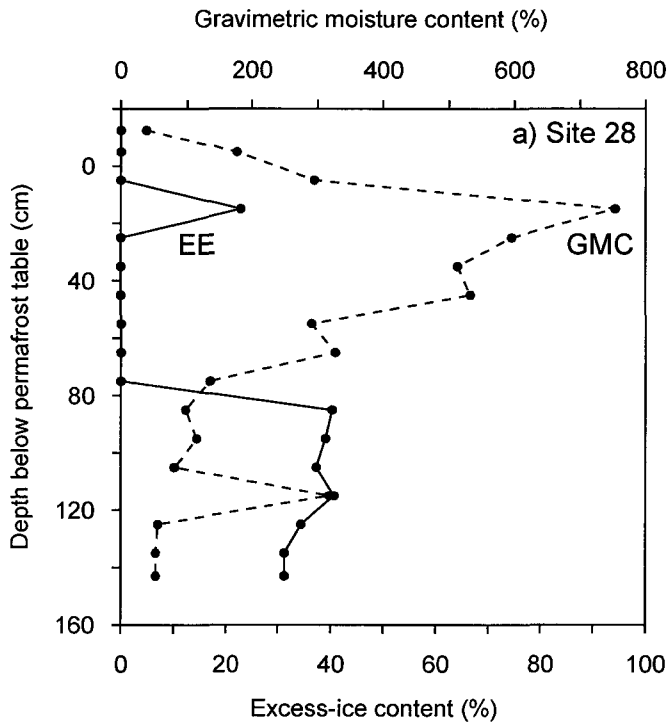
The three-layer, qualitative model of the active layer-permafrost system presented by Shur et al. (2005) is helpful to conceptualize and understand the dynamics of frozen ground. However, results from this study highlight some challenges in applying the classification to data collected in the field. The qualitative nature of the model makes it difficult to identify or measure the boundaries between the transient layer (the uppermost part of the transition zone) and the rest of the transition zone.

Shur et al. (2005, p. 7) indicated that the transient layer 'contains less ice than underlying regions because it undergoes thaw more frequently'. In order to identify this

layer, either the ice contents of near-surface permafrost must be examined, or active-layer depth records must be analyzed to determine frequencies of thaw at different depths. Assessment of the transient layer based on ice content may prove difficult, because ice characteristics of a core may not be apparent during sampling. In addition, the location of the transient-layer boundary must be identified arbitrarily from ice content profiles, unless a quantitative definition of the boundary is available.

The distinction between the transient layer and the transition zone may be complicated by the organic-rich layer commonly found at the top of permafrost in hummocky tundra terrain. The results from this study have shown that there may be no measureable excess ice in organic-rich core segments despite high volumes of ice, making it difficult to determine whether lower ice contents are a result of periodic thaw of the transient layer or soil organic characteristics. For example, Site 28 in the tundra had 79% organic matter content between 10 and 20 cm below the 2010 permafrost table, and an excess-ice content of 22% (Figure 5.2a). The top 10 cm of permafrost contained no excess ice. Between 20 and 50 cm below the permafrost table, the volume of water in core segments was similar to that between 10 and 20 cm, but there was no excess-ice, likely due to high organic-matter contents observed at these depths. Beyond 80 cm below the permafrost table, the organic matter content was lower and excess-ice was measured in all core segments. These variable soil conditions in the tundra may complicate the identification of the transient layer based on the ice contents of upper permafrost.

Even for cores with little organic matter, identifying the transient layer may prove difficult. For example, Site 35 in the tundra had 4% organic matter in the top 10 cm of permafrost, where the soil was 78% silt and 22% clay. The excess-ice and gravimetric



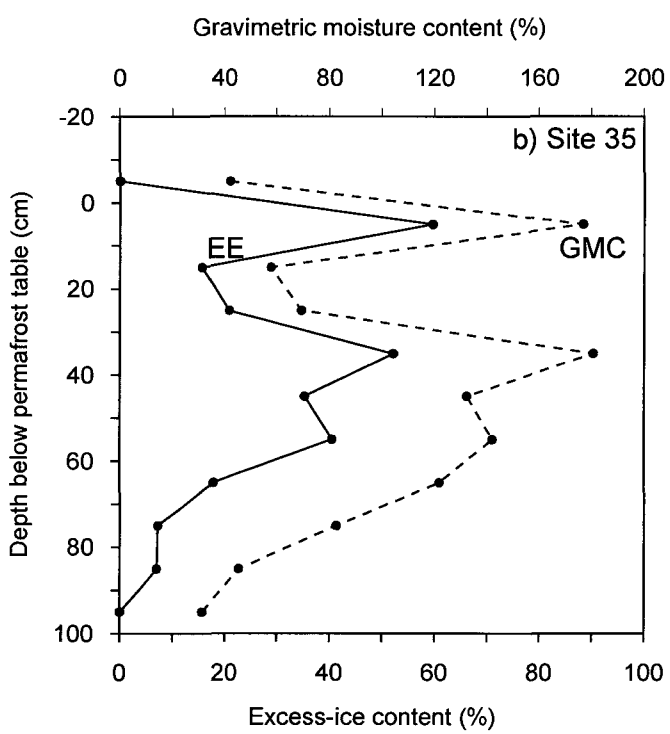


Figure 5.2 Gravimetric moisture and exess-ice content profiles of (a) Site 11, an organic rich site in the tundra, and (b) Site 51, an organic poor site in the tundra. The solid line is the excess-ice content and the dashed line is the gravimetric moisture content.

moisture contents were highest in this core segment immediately below the 2010 active layer depth (Figure 5.2). The excess-ice content declined between 10 and 30 cm below the permafrost table, and then rose again between 30 and 40 cm. Below these depths, the excess-ice and gravimetric moisture contents generally declined. In this case, the highest ice content was at the very top of permafrost, where the periodic thaw of the transient layer should limit the ice content. One possible explanation for this discrepancy may be that the transient layer at this location is very thin, and cannot be resolved by drilling 10 cm core segments.

Identification of the transient layer based on a qualitative definition of the relative ice content is difficult, particularly in the heterogeneous soil and organic conditions present in hummocky tundra terrain. Shur et al. (2005) indicated that the transient layer is not typically a subject of research in active-layer studies, because long-term thaw depth records are largely unavailable. The authors suggested that examination of active-layer thicknesses over a minimum of three decades would likely be necessary to capture site-specific frequency-magnitude patterns necessary to evaluate the transient layer. Active-layer records of this length will shortly be available from the tundra near Illisarvik, and future research could then attempt to characterize the transient layer based these data.

6. SUMMARY AND CONCLUSIONS

6.1 Summary of results

This thesis has examined near-surface ground ice development in the Illisarvik drained lake basin and the surrounding tundra, and investigated relations between controlling factors on ice accumulation. In addition, active-layer development over the past three decades was examined. Near-surface ground ice contents were highly variable in the basin, and ice enrichment was observed to a maximum depth of 60 cm below the 2010 permafrost table. Active-layer soil moisture status appeared to be the primary control on ice accumulation. There was much more near-surface ground ice in permafrost at tundra sites than in the basin, likely due to aggradation of the permafrost table since the early-Holocene climatic optimum and ice enrichment from summer moisture migration to the top of permafrost. Active-layer variations in the basin and tundra were significantly associated with summer temperatures, and in some basin areas, to snow depths. Correlations between air temperatures and active-layer thicknesses in the basin have decreased since drainage, concurrent with vegetation establishment and surface organic matter accumulation. Active-layer variations on the lake basin terrace were insensitive to air temperatures and snow depths, likely due to the thick layer of peat that covered the surface.

6.1.1 Near-surface ground ice development in the lake basin, tundra and lake terrace

Wet willow (WW) areas in the lake basin had the greatest near-surface ground ice contents, averaging 10% in the top 50 cm of permafrost. Sites near the lake margin and

with little vegetation (LM) had a mean excess-ice content of 2% in the upper 50 cm of permafrost, the least of all basin areas. Sites in the basin centre (BC) had an intermediate mean excess-ice content of 4% in the top 50 cm of permafrost.

Active-layer soil moisture status appeared to be the dominant control on ice development in the lake basin. Despite similar soil conditions and active-layer histories, sites in WW had substantially more near-surface ground ice than sites in BC. In addition, drill sites at thirteen wet locations in the lake basin had statistically greater ice contents than thirteen sites in dry locations.

Silt contents at the top of permafrost in the lake basin were between 16 and 83%, with a median value of 74%. Excess-ice contents at the top of permafrost were generally higher at sites with higher silt contents, and no excess-ice was present at sites with more than 55% sand content. However, the large variation in ice contents observed at sites with high silt contents suggests that soil texture was a secondary control of ice accumulation.

Organic matter content at the top of permafrost in the basin ranged between 2 and 25%, and was not associated with excess-ice content of upper permafrost. There was a significant statistical relation between the organic matter content of upper permafrost and the distance from the former lake shore, likely because, before drainage, vegetation was undisturbed below the maximum depth of winter ice cover in deeper areas near the lake centre.

There was no statistical relation between vegetation height and ice accumulation. It was hypothesized that sites with higher vegetation would have warmer permafrost due to increased snow entrapment, and that upward winter moisture migration from the top of

permafrost would be greater at sites with warmer permafrost due to higher hydraulic conductivities at the prevailing temperatures. However, either this is not physically important or the small number of annual cycles since lake drainage has limited the influence of permafrost temperatures on seasonal moisture migration to and from the top of permafrost.

Active-layer history in the basin appears to have a limited influence on near-surface ground ice formation. The greatest aggradation of the permafrost table occurred in BC sites, but the ice contents were much lower than in WW. The amount of permafrost aggradation in the basin since 1979 is small compared to aggradation in the tundra since the early Holocene, and appears of little importance to the quantity of near-surface ground ice in comparison with soil moisture and texture.

In the tundra, the mean excess-ice content of the top 50 cm of permafrost was 25%, much higher than at basin sites. Near-surface excess-ice was also present at greater depths in the tundra, to 1.5 m below the 2010 permafrost table. The average excess-ice content of the top metre of permafrost was 20%. The high near-surface ground ice contents in the tundra are likely a result of aggradational ice formation as the permafrost table rose following the early-Holocene climatic optimum, with thousands of seasonal cycles allowing a net moisture migration to the top of permafrost to form repeated-segregation ice.

The mean excess-ice content of the top 50 cm of permafrost on the lake terrace was 6%. The majority of ice enrichment at terrace sites occurred at greater depths, and the mean excess-ice of the top metre of permafrost was 9%. Excess-ice was not present in

the top 20 cm of permafrost at terrace sites, because the peat layer on the terrace that extended into permafrost was able to accommodate all liquid water when icy cores were thawed.

6.1.2 Active-layer development

Basin active layers have thinned since the 1990s in WW and BC. In LM, mean active-layer depths have remained similar in the three decades since drainage. Active layer depths in BC and LM were strongly and significantly correlated with summer temperatures in the 1980s. These correlations weakened and became insignificant, concurrent with establishing vegetation in the 1990s and 2000s, suggesting that after drainage, basin permafrost was initially climate-driven and then changed into climate-driven, ecosystem modified permafrost. Correlations between active-layer depths and summer temperatures were low and insignificant in WW in the 1980s, where vegetation establishment was most rapid.

Summer air temperatures and snow depths, in WW and BC only, explained between 31 and 61% of active-layer variation in the basin. The low r^2 values in WW and LM suggest that the basin active-layer system is complex in these areas and responds to a number of variables, including site-specific soil moisture and vegetation characteristics that were not measured in this study.

Tundra hillslope active-layer depths have remained unchanged in the past three decades, and they have been increasing slowly on the tundra hilltop. Linear regressions indicated that summer air temperatures explained 55% of the variation in active-layer depths at hillslope sites and 36% at hilltop sites, suggesting that variables not measured in

this study are important to tundra active-layer depths. Variation in snow depths was likely not sufficient to significantly influence active-layer depths in the tundra.

There was no significant trend in active-layer depths on the terrace, and active-layer depths were insensitive to seasonal air temperatures and snow depths. The thick layer of peat and its influence on thermal properties is likely responsible for the unique active-layer behaviour on the terrace.

6.1.3 Snow depths

Snow depths in the basin were significantly correlated with vegetation height. In the basin and tundra, snow depths have been increasing significantly since the 1980s, but have remained constant at Benchmark 41 on the lake basin terrace, where vegetation is minimal. These results suggest that vegetation growth may be responsible for increasing snow depths in the basin and tundra.

6.2 Conclusions

The following five conclusions can be drawn from the examination of near-surface ground ice development, controlling factors on ice accumulation, and active-layer history at the study sites:

(1) Near-surface ground ice development was observed up to a depth of 60 cm below the permafrost table in the lake basin, and was controlled mainly by soil moisture conditions in the active layer, and secondarily by soil texture.

(2) Near-surface ground ice contents in the tundra were much greater than in the basin, and ice was found at much deeper depths, likely due to aggradational and repeated-

segregation ice formed as the permafrost table aggraded since the early-Holocene climatic optimum.

(3) Basin active layers are complex systems that are sensitive to a number of variables, and were associated predominantly with summer air temperatures and in some areas, snow depths. Active-layer thinning was less important than other factors for excess-ice accumulation in the basin. Active layers in the tundra also respond to a number of variables and were associated with summer air temperatures.

(4) Lake terrace active-layers were insensitive to air temperatures and snow depths, probably due to the thermal properties of the peat layer and the latent heat required to thaw the abundant ice observed in terrace peat cores.

(5) Snow depths in the lake basin and adjacent tundra have been increasing, most likely due to vegetation growth.

6.3 Research implications and direction for future studies

The results of this research have implications related to resource management, climate change, and future ground ice studies. Resource managers and industry should be aware of variable near-surface ground ice conditions that may occur in different terrain types, in order to plan accordingly for construction projects. The importance of soil moisture, texture, and active-layer variations over long periods should be considered due to their influences on near-surface ground ice accumulation.

If air temperatures and snow depths in the tundra continue to increase, the permafrost may warm and thaw near-surface ground ice. This may increase the frequency

of active-layer detachment slides, and certainly cause ground subsidence, both posing potential hazards to existing infrastructure.

Valuable future research may further investigate near-surface ground ice and controlling factors in different study areas and landscape types. In addition, future studies at Illisarvik may utilize the results from this thesis to monitor longer-term ice development in the lake basin and tundra. Such investigations may observe how changing vegetation and climate affects near-surface ground ice.

Future research may further examine active-layer variations at the study site. The thirty year active-layer record from the tundra transect may enable quantification and examination of the transient layer.

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