NEAR-SURFACE GROUND ICE IN SEDIMENTS OF THE MACKENZIE
DELTA REGION, NORTHWEST TERRITORIES

by

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A thesis submitted to
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in partial fulfillment of
the requirements for the degree of

Doctor of Philosophy.

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“NEAR-SURFACE GROUND ICE IN SEDIMENTS OF THE MACKENZIE
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Abstract

Physical and chemical characteristics of near-surface permafrost in the Mackenzie Delta region, N.W.T., were studied in the summers of 1999 to 2001. In sediments of the Mackenzie Delta, ice-poor permafrost was encountered on aggrading point bars colonized by willows. A near-surface ice-rich zone, 1 to 2 m thick, occurred in silt loam soils at spruce forest sites elevated above the level of annual flooding. Lower-lying mesic sites colonized by alder and sedge communities were also underlain by ice-rich permafrost. The factors that influence ground ice accumulation in sediments of Mackenzie Delta are: (1) sediment texture; (2) sedimentation rate; (3) soil moisture; and (4) ground-thermal regime.

In hummocky terrain adjacent to the Delta, ice-poor permafrost developed beneath a rapidly aggrading permafrost table (4 cm a^{-1}), but a near-surface ice-rich interval 10 to 20 cm thick developed where the permafrost table remained static for over two decades. The maximum thickness of the ice-rich intervals beneath decadal and millennial thaw unconformities was 30 cm, indicating that near-surface ground ice may accumulate in the absence of an aggrading permafrost table. Under a static permafrost table, initial rates of ice accumulation were on the order of 1 to 10 mm a^{-1} , but total ice accumulation above a millennial thaw unconformity indicates that in aggregate, these initial rates must decrease with time.

Permafrost was nutrient-rich relative to the mineral soils at the base of the active layer throughout the study region. Nutrient enrichment of permafrost with respect to overlying active-layer sediments was greatest at ice-rich sites. Within a profile,

the correspondence between ice-rich intervals and zones of cation enrichment indicates that solutes migrate with water along thermally-induced suction gradients and accumulate in the near-surface ice-rich zone.

There is a clear, positive association between the ice content of near-surface permafrost and the extent of tree lean at sites in the Mackenzie Delta, suggesting that aggradation and/or degradation of ice-rich permafrost may cause spruce trees to tilt. The effect of ground-ice development on the tilt of spruce trees was corroborated at a black spruce site in hummocky terrain where reaction-wood rings indicating tree tilting was produced during periods of active-layer thinning and aggradational ice development.

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Frontispiece. Oblique aerial photograph of a point bar in central Mackenzie Delta. June 22, 2001.

CHAPTER 1. INTRODUCTION

1.1 Introduction

In permafrost terrain, an ice-rich zone of segregated origin is frequently encountered beneath the base of the active layer (Pollard and French, 1980; Burn, 1988; Shur, 1988). This ice-rich zone is of significance because: (1) it may act as an important hydrological reservoir (Hughes, 1974); (2) its thawing can result in local subsidence and terrain instability (Mackay, 1970); and (3) it may act as a thermal buffer to ground warming due to the high latent heat content of ice (Smith, 1986). Furthermore, the development of near-surface ice-rich permafrost may influence surface micro-relief in hummocky terrain (Mackay, 1995).

This thesis investigates the physical and chemical characteristics of the near-surface ice-rich zone in the Mackenzie Delta region (Figure 1.1). Specifically, the objectives of this research on near-surface ground ice are to: (1) measure its spatial variability with respect to environmental conditions in the Mackenzie Delta; (2) investigate the vertical extent of moisture accumulation; (3) investigate its development through time; (4) describe its geochemical characteristics; and (5) examine the effect of aggradational ice development and degradation on spruce forest structure.

The study includes the description and comparison of physical properties of near-surface permafrost of different ages, and of permafrost that has developed under various environmental conditions. The geochemical properties of near-surface permafrost from various geomorphic environments are described and the process of

Figure 1.1. Map of the Mackenzie Delta region.

nutrient accumulation in near-surface permafrost is discussed. The effect of ground-ice aggradation and degradation on spruce forest structure was investigated at sites in the Mackenzie Delta and on hummocky terrain near Inuvik.

The Delta environment facilitated investigation of the physical and chemical evolution of the near-surface ice-rich zone. Core samples obtained from transects perpendicular to stream channels described permafrost conditions across environmental gradients. Historical active-layer data collected by Mackay (1995) at a site adjacent to the Mackenzie Delta near Inuvik provided a context within which to interpret vertical patterns of ice-enrichment at that site, and enabled the estimation of ground-ice accumulation over annual and decadal scales. The effect of aggradational ice development on microtopography and forest structure was also investigated at these sites.

1.2 Permafrost and the active layer

Permafrost is defined as ground (soil or rock) that remains at or below 0°C for at least two years (Associate Committee on Geotechnical Research, 1988). The critical variable in determining the distribution of permafrost is ground-surface temperature which is a function of the surface energy balance (Oke, 1978).

Permafrost distribution is not only dependent on regional climate, local factors such as snow depth, surface wetness and the nature of vegetation cover commonly override the influence of larger scale macroclimatic factors on ground-thermal conditions (Smith, 1975; Smith and Riseborough, 1983).

The active layer is the top layer of ground subject to annual thawing and freezing in areas underlain by permafrost (Associate Committee on Geotechnical Research, 1988). Active-layer thickness exhibits a significant amount of spatial variability that reflects summer surface temperature (Mackay, 1970), thickness of surface organic materials (Zoltai and Tarnocai, 1975), vegetation cover (Mackay, 1995), water content of soil materials (Farouki, 1981) and depth of winter snow cover (Smith, 1975).

Climate variability, forest fires and vegetation succession influence temporal variability in active-layer depths (Mackay, 1995; Burn, 1997). A widespread thaw unconformity that indicates maximum extent of active-layer development during the early Holocene warm interval has been reported in central Yukon and western Arctic Canada (Mackay, 1975; Burn *et al.*, 1986; Murton and French, 1993; Burn, 1997; Kokelj *et al.*, 2002).

1.3 Water movement in frozen soils

The development of the near-surface ice-rich zone is linked to the ground-thermal regime and to the behaviour of water in freezing and frozen soils (e.g., Burt and Williams, 1976; Horiguchi and Miller, 1983; Mackay, 1983). It has been known for some time that not all water freezes at 0°C (Nersesova, 1953), and laboratory experiments have shown that water can move through frozen soil under the influence of both hydraulic (Burt and Williams, 1976) and temperature-induced suction gradients (Horiguchi and Miller, 1983). However, the unfrozen moisture

content and permeability of frozen soils decrease by several orders of magnitude at temperatures just below 0°C. Several field studies have identified temperature-induced migration of water in frozen soil (e.g., Parmuzina, 1978; Mackay, 1983; Burn, 1988).

1.4 Solute redistribution during ground freezing

Hypotheses designed to examine the chemical characteristics of the ice-rich zone have been based upon a theoretical and laboratory understanding of solute movement with respect to moisture during the freezing of soils (e.g., Hallet, 1978; Qui *et al.*, 1988; Chuvilin *et al.*, 1998). During the freezing of coarse-grained sediments, exclusion of salt from the ice phase redistributes solutes into the unfrozen soil ahead of the advancing freezing front (Hallet, 1978; Mackay, 1997, 1998). In contrast, laboratory experiments conducted with fine-grained silt and clay have demonstrated that there is a net movement of moisture and ions towards the freezing zone (Cary and Mayland, 1972; Qui *et al.*, 1988; Chuvilin *et al.*, 1998). Qui *et al.* (1988) concluded that in most silt and clay, ion redistribution due to moisture migration and temperature diffusion towards the freezing front will greatly exceed the amount of concentration diffusion away from the frozen zone.

1.5 Ground ice

Ground ice is broadly defined as moisture frozen beneath the ground surface (Harry, 1988). It may be preserved indefinitely by the presence of permafrost. The

physical significance of permafrost is derived from the presence of excess ice, which refers to the volume of supernatant water present if a vertical column of frozen sediment were thawed (French, 1996). Thawing of permafrost containing excess ice will result in ground subsidence, and on slopes, thawing of ice-rich soil may induce failure along the surface of permafrost (Mackay, 1970; Lewkowicz, 1990).

In permafrost terrain, ground ice can be classified as buried ice or intrasedimental ice (French and Harry, 1990). Buried ice has been identified in the Mackenzie Delta region (Dallimore and Wolfe, 1988; French and Harry, 1990), but most ground ice in this region is intrasedimental ice. Intrasedimental ice includes water that reaches the permafrost by gravity transfer and freezes to form wedge (Mackay, 1972), vein (Gell, 1978), dilation crack (Mackay, 1985) or pool ice (Mackay, 1997). Also, water may freeze *in situ* to form pore ice, or may be injected under pressure into mineral or organic deposits, forming layers of intrusive ice (Mackay, 1972). In freezing and frozen soils, moisture migration driven by thermally-induced suction gradients can form thick segregated ice lenses (Mackay, 1971, 1972).

1.6 Aggradational ice

Permafrost at the base of the active layer is characteristically ice-rich (Figure 1.2) (Mackay, 1972; Burn, 1988). Mackay (1972) attributes the presence of the ice-rich zone to a rising permafrost table as the result of climate change, vegetation succession or sedimentation, trapping ice lenses formerly at the base of the active layer. Cheng (1983) suggested that the upper part of permafrost is a net sink for active-layer water.

Figure 1.2 Typical distribution of ground ice with depth, Richards Island, NWT
(Pollard and French, 1980, Figure 2, Data from Lawrence and Proudfoot, 1977).

In summer, active-layer soil water is drawn downward along a thermally-induced suction gradient into permafrost. The downward moisture flux in summer should be greater than water movement out of permafrost in winter, when the temperature gradient is reversed (Cheng, 1983). The flux imbalance can be attributed to seasonal change in the hydraulic conductivity of near-surface permafrost soils, which are several orders of magnitude lower in the winter when soil temperatures are much colder than in summer.

Downward migration of water into permafrost is supported by seasonal observations of moisture migration and frost heave (Parmuzina, 1978; Mackay, 1983; Smith, 1985; Burn and Smith, 1985) and through analysis of atmospherically-derived tritium within the permafrost profile (Michel and Fritz, 1978, 1982; Chizhov *et al.*, 1983; Burn and Michel, 1988). The development of aggradational ice can result in surface heaving and an increased hummocky micro-relief, whereas active-layer deepening and thawing of ice-rich permafrost can cause surface subsidence (Mackay, 1970, 1995).

1.7 Solute accumulation in the near-surface ice-rich zone

Laboratory experiments involving the freezing of fine-grained silt and clay have demonstrated the coupled movement of water and soluble ions (Cary and Mayland, 1972), but solute redistribution has not been considered in the context of aggradational ice development. The zones of solute accumulation in freezing fine-grained soil columns generally coincide with ice enrichment (Cary and Mayland,

1972; Qui *et al.*, 1988; Chuvilin *et al.*, 1998). In permafrost terrain underlain by fine-grained silt and clay, ions may be drawn to the top and bottom of the active layer during winter freeze-back as moisture migrates towards the descending and ascending freezing fronts (Mackay, 1983). The results of laboratory experiments (Cary and Mayland, 1972; Baker and Osterkamp, 1988; Qui *et al.*, 1988; Chuvilin *et al.*, 1998) suggest that solute movement should accompany downward moisture migration into the top of permafrost, and, thus, the near-surface ice-rich zone should be solute-rich.

1.8 Study area

This study was conducted in the eastern part of the Mackenzie Delta and on terrain adjacent to the Delta near Inuvik. The modern Mackenzie Delta is approximately 150 km long and 60 km wide and consists of alluvium deposited by the Mackenzie and Peel rivers (Figure 1.1). The Delta is characterized by an intricate network of migrating channels and numerous lakes (Mackay, 1963). Channel migration results in a complex interaction between topography, vegetation, microclimate and thermal regime (Gill, 1973; Smith, 1975). Vegetation along point bars displays a sequential pattern which, through trapping of snow and development of an organic horizon, affects ground temperatures and active-layer thickness (Smith, 1975). Permafrost thickness increases with distance away from the migrating channel as the degree of thermal recovery increases with age of the land

surface (Smith, 1976). Permafrost is generally less than 100 m thick due to the thermal effect of water bodies and the shifting nature of the channels (Smith, 1976).

Wedge ice is present throughout sediments of the Mackenzie Delta and pingos occur in low-lying terrain of the outer Delta (Mackay, 1963). Ground-ice investigations near Aklavik conducted in the 1950's indicated that segregated ice lenses were concentrated in the top 1 to 3 m of Delta alluvium (Pihlainen and Johnston, 1954), but the distribution of near-surface ground ice in sediments of the Delta is not well understood.

Study sites adjacent to the Mackenzie Delta near Inuvik were in hummocky terrain vegetated by black spruce (*Picea mariana*). The Mackenzie River does not flood terrain at these sites. Permafrost thickness is approximately 90 m (C.R. Burn and L. Kutny, pers. comm., 2002) and the near-surface sediments, predominately clayey silts, are ice rich (Watson *et al.*, 1973).

1.9 Objectives and organization

This thesis builds on work by Cheng (1983) and Mackay (1983) who proposed that ice accumulation in near-surface permafrost may be part of the annual water balance of permafrost terrain, and by Burn (1986) who investigated the mechanisms determining the water balance of permafrost.

The objectives of this research may be summarized as:

- 1) To examine the spatial distribution of ground ice in sediments of the Mackenzie Delta;

- 2) To investigate the vertical extent of moisture movement into permafrost in the Mackenzie Delta and at Navy Road;
- 3) To estimate rates of ground-ice accumulation over different time scales;
- 4) To determine if the near-surface ice-rich zone of permafrost is a sink for soluble nutrients.
- 5) To examine the effects of aggradational ice development and degradation on spruce forest structure;

The thesis is arranged into nine chapters. The following chapter (Chapter 2) provides a review of literature pertaining to the movement of water and soluble ions in frozen soils and establishes a theoretical context for this study. An overview of the study region is then provided (Chapter 3). Research hypotheses and research design are summarized and study sites are described in Chapter 4. The field and laboratory methods are also described in this chapter. Results are divided into four chapters.

The focus of Chapter 5 is near-surface ground-ice characteristics in the Mackenzie Delta. Ground-ice characteristics were described across a range of environmental conditions encountered in the Mackenzie Delta, enabling factors that control aggradational-ice development to be examined. The change through time of the physical characteristics of the ice-rich zone was investigated in the Mackenzie Delta where the relative age of permafrost increases with distance from shifting channels (Smith, 1976). In Chapter 6, the vertical distribution of moisture content patterns in hummocky terrain at Navy Road are examined in the context of a 30-year record of active-layer history at that site (Mackay, 1995). The decadal record of permafrost

table fluctuations at this site in conjunction with perturbation experiments enabled investigation of ice accumulation rates over various temporal scales. In Chapter 7, the geochemical characteristics of near-surface permafrost are described for a range of environments throughout the study area, and soluble ion profiles in near-surface permafrost are described and compared with patterns of moisture accumulation in the ice-rich zone of permafrost. The effect of ground-ice development and degradation on forest structure is examined in Chapter 8. Conclusions are summarized in Chapter 9.

CHAPTER 2. REVIEW OF LITERATURE: HYDROLOGIC AND GEOCHEMICAL PROCESSES IN NEAR-SURFACE PERMAFROST

2.1 Permafrost hydrology

Investigations of surface hydrology in permafrost terrain consider the permafrost table as an aquiclude (Woo and Steer, 1986; Woo and Zia, 1995). This characteristic is conducive to the swift removal of snowmelt or rainfall from permafrost terrain, resulting in rapid catchment responses to hydrological inputs (Cogley and McCann, 1976). Although generally not considered in water balance calculations, some surface water may infiltrate into frozen soils (Kane, 1980; Marsh, 1988).

Most of the water movement within the thawed active layer is driven by gravity (Woo and Zia, 1995). This assumption is valid during spring melt when the majority of water is removed from slopes by surface and subsurface flow (Lewkowicz and French, 1982; Kokelj and Lewkowicz, 1998). Hydrologic processes within the active layer, unique to regions underlain by permafrost, occur at the base of the active layer during the thaw season and within the active layer during freeze-back (Parmuzina, 1978; Cheng, 1983; Mackay, 1983). Although several orders of magnitude less than in unfrozen soils, frozen soils have a measurable permeability (Burt and Williams, 1976), and moisture movement is controlled by strong suctions induced by temperature gradients rather than by gravity (Williams and Smith, 1989).

The hydrology of frozen ground is closely linked with the ground thermal regime (Williams and Smith, 1989). The behaviour of water in freezing soils is fundamental to explaining the physical and chemical evolution of the near-surface ice-rich zone of permafrost. The first part of this chapter examines the behaviour of water in frozen soils. The remainder of the chapter describes the annual hydrological regime of permafrost and the physical and chemical evolution of the ice-rich zone is examined.

2.2 Hydrology of frozen soils

Considerable proportions of water in soils just below 0°C remain in the liquid form (Burt and Williams, 1976). In frozen ground, soil water is found in small capillaries and as thin films adsorbed onto soil particles (Williams and Smith, 1989). The unfrozen water content of a soil declines with temperature and is described by an unfrozen moisture content characteristic curve, each unique to a specific soil (Figure 2.1) (Burt and Williams, 1976).

The presence of unfrozen water is due to several factors. Most important are the electrochemical effects in pore water associated with the electrical fields emanating from soil-particle surfaces that act to prevent water molecules from becoming a part of the crystal lattice of ice (Mitchell, 1976). Solutes excluded from the ice phase during the formation of pore ice will increase dissolved ion concentrations in the unfrozen water (Hallet, 1978), further depressing the freezing temperature of water (Banin and Anderson, 1974). As temperatures continue to drop and ice grows

Figure 2.1 Unfrozen water content characteristic curves for a range of fine-grained soils (Burt and Williams, 1976, Figure 1).

within the soil matrix, water is confined to smaller pores, decreasing the ice nucleation point (Everett, 1961).

The quantity of unfrozen water present in a soil at a given temperature is related to the soil texture and mineralogy (Burt and Williams, 1976) (Figure 2.1). Clays with a high specific surface area and strong electro-negative charges are able to hold a considerable amount of water at very high tension, allowing water to exist in equilibrium with ice at temperatures well below 0°C. Based on this soil property, the clay content of a soil may be used as an indication of unfrozen water content (Anderson and Tice, 1972).

2.3 Water potential gradients due to soil freezing

In fine-grained freezing soils, the potential difference between ice (high pressure) and the adjacent water (low pressure) is proportional to the temperature depression below 0°C (Williams and Smith, 1989). The Clausius-Clapeyron relationship describes the strong suction (dP_w) which develops as soil freezing progresses and temperature falls (Williams, 1982).

$$[2.1] \quad dP_w = -\Delta T L / VT$$

where:

ΔT is the lowering of the freezing point (K);

L is the latent heat of fusion ($J \text{ kg}^{-1}$);

V is the specific volume of water ($m^3 \text{ kg}^{-1}$);

T is temperature (K).

Equation (2.1) indicates that a lowering of the temperature by 1°C induces a suction of 1.2 MPa (Williams, 1982). Thus, a temperature gradient in frozen fine-grained soils will induce heat and water to flow in the same direction.

The behaviour of water during freezing of coarse-grained materials differs from that in silt and clays. Sands have a low specific surface area and sandy soils have larger pore spaces than silt and clay soils. In sands, much of the soil water is contained within large pore spaces and is not at high tension, freezing at temperatures just below 0°C (Burt and Williams, 1976). The low hydraulic conductivity of frozen sands (Burt and Williams, 1976) inhibits moisture migration along temperature-induced suction gradients. Instead, the 9% expansion of freezing pore-water in sands can squeeze unfrozen water out from soil pores in the direction of lower water pressure (Mackay, 1997, 1998).

2.4 Moisture transfer mechanisms in frozen soils

Laboratory experiments indicate that hydraulic conductivity of frozen soil is low, except at temperatures just below 0°C (Figure 2.2). However, soils used in the laboratory determination of hydraulic conductivity are subject to two major deviations from field conditions (Mackay, 1983). Natural soils are heterogeneous in texture and thus, unfrozen moisture content and hydraulic conductivity may exhibit considerable spatial variability. Secondly, experimental work has suggested that freeze-thaw cycles can greatly increase the hydraulic conductivity of a remolded soil (Chamberlain and Gow, 1979).

Figure 2.2 Hydraulic conductivities of fine-grained materials. Chena and NWA are natural silts from Fairbanks, Alaska, Manchester is a silt from New Hampshire and Calgary is a silt obtained near Calgary, Alberta. The sample designated '4-8 μm Fraction' was separated from the Manchester sample. Specific surfaces of each sample are given. Data measured during chilling are plotted as open symbols and data measured during warming are plotted as solid symbols (Horiguchi and Miller, 1983, Figure 2).

Water can also migrate through ice lenses in frozen soils via a translational movement termed “regelation” (Miller, 1970; Burt and Williams, 1976). Miller (1970) used an “ice- sandwich” permeameter to demonstrate that ice lenses do not inhibit water movement through frozen soils. Observations of secondary ice lens growth in freezing soil columns by Ohrai and Yamamoto (1985) support Miller’s observations.

Regelation transport can be explained in terms of pressure differences between water and ice on the warm and cold sides of an ice lens. According to the Clausius-Clapeyron equation, if the freezing point is less than 0°C, the pressure of ice will exceed the pressure of water (Williams and Smith, 1989). As water moves in the direction of a thermally-induced pressure gradient, the pressure of water at the water/ice interface will rise (pore-water tension will decrease) and the equilibrium freezing temperature will rise above ambient. As a result, freezing can occur at the warm side of the ice lens. The growth of the lens will increase its pressure, which will be reflected on the cold side of the lens by an increase in the difference between the pressure of ice and the pressure of water, inducing melting. Ice melted at the cold side of the ice lens can proceed to migrate in liquid phase along the thermally-induced pressure gradient to a second ice lens well behind the primary freezing front (Ohrai and Yamamoto, 1985).

2.5 The relative importance of water movement in different phases

The relative importance of transport mechanisms in frozen soil can be expected to vary with the total moisture content of soil and temperature (Williams and Smith, 1989). In saturated soils, Burt and Williams (1976) found that ice lenses reduced values of hydraulic conductivity in fine-grained soils at temperatures very close to 0 °C. Hydraulic conductivities for fine-grained soils at temperatures above -0.1°C (10^{-8} and $10^{-11} \text{ m s}^{-1}$) measured by Horiguchi and Miller (1983) are greater than the transport coefficient of ice at similar temperatures given by Wood and Williams (1985) (Figure 2.3). However, at colder temperatures, conductivity drops rapidly (Figure 2.2) and is approximately two orders of magnitude lower than the transport coefficient for the ice phase. This suggests that at temperatures below about -0.1°C, the limiting factor determining water movement through frozen soil is the rate at which unfrozen water can arrive at the warm side of an ice lens in order to sustain the regelation process (Williams and Smith, 1989).

In unsaturated soils, moisture moves through pore spaces by vapour diffusion and in adsorbed films of water by liquid flow. Using Darcy's equation for liquid flow rate and vapour flow, Williams and Smith (1989) illustrated that for given soil temperature conditions, liquid flow is several orders of magnitude greater than vapour flow. As water content increases, the relative importance of vapour diffusion drops rapidly because potential paths for vapour transport are filled by pore ice (Williams and Smith, 1989). Although vapour flow may be an important mechanism of heat and moisture transfer within an under-saturated active layer

Figure 2.3 Relationship between temperature and the transport coefficient of ice (Wood and Williams, 1985, Figure 3).

(Hinkel and Outcalt, 1994), it is of low relative importance in the near-surface ice-rich zone.

2.6 Moisture movement through frozen soils and ice segregation

In frozen soils, water moves in the direction of a positive temperature gradient under the influence of a thermally-induced pressure gradient (see Section 2.3). High suctions develop in a zone of partially frozen soil behind the freezing front known as the frozen fringe (O'Neill and Miller, 1985). Suction gradients result from temperature-induced changes in the unfrozen moisture content and pore water pressures within the frozen fringe. A sharp decrease in conductivity between frozen and unfrozen soil results in a flux divergence causing moisture to accumulate at the point of maximum flux divergence at some distance on the cold side of the 0° isotherm (Williams and Smith, 1989). This accumulation increases the pressure of water, producing conditions that favour freezing and growth of an ice lens (Williams and Smith, 1989). The freezing of water releases latent heat that must be dissipated for the temperature to remain constant and for the ice lens to continue to grow. If suction is maintained and water can be continually supplied through the frozen fringe, the ice lens will grow until the rate of heat extraction exceeds the amount of latent heat released by the freezing of water at the ice lens (Williams and Smith, 1989). Soil cooling will decrease the hydraulic conductivity within the frozen fringe (Horiguchi and Miller, 1983), impeding the supply of water to the

growing lens. Since heat is continually being extracted from this system, the freezing front will rapidly advance to a new position of maximum flux divergence.

Although it has been suggested that ice lenses may inhibit moisture movement through frozen soils at temperatures just below freezing (Williams and Smith, 1989), Ohrai and Yamamoto (1985) observed that several ice lenses grew simultaneously over distances of a few centimetres in a freezing column of fine-grained saturated clay. Existing ice lenses should be preferred locations for continued ice enrichment because at colder temperatures, regelation transport is limited by the hydraulic conductivity of frozen soil on the cold side of the ice lens (Williams and Smith, 1989).

2.7 The annual water balance of permafrost

The seasonal sequence of water movement within the active layer and the top of permafrost is examined in this section (Figure 2.4). The hydrological year is divided up into spring snowmelt, summer thaw, fall freeze-back and winter.

During spring snowmelt, latent heat is released by the refreezing of meltwater at the base of the snowpack, raising near-surface soil temperature and establishing conditions favourable to temperature-induced movement of surface water into the underlying frozen soil (Kane, 1980; Marsh, 1988). Some meltwater may move down desiccation cracks in the frozen soil, but these pathways are rapidly blocked by ice (Mackay, 1984). Continued warming of frozen soils beneath the snowpack is thought to be the result of infiltration and refreezing of meltwater (Marsh, 1988).

Figure 2.4 Vertical water movement in the active layer and near-surface permafrost (Mackay, 1983, Figure 8).

Moisture movement into dry frozen alluvium may account for significant storage of meltwater during spring melt (Marsh, 1988).

Active-layer thaw commences once the ground is free of snow and ice.

Thawing is accompanied by downward moisture migration from thawed soil into the warming frozen ground below (Figure 2.4) (Parmuzina, 1978; Mackay, 1983; Smith, 1985). Parmuzina (1978) reported an increase of 10 to 25% in gravimetric moisture content of the frozen active layer during summer thaw and observed a similar increase in the upper part of permafrost. Studies by Cheng (1983) reported summer increases in the water content of near-surface permafrost of up to 0.05 g cm^{-3} (Figure 2.5). Moisture redistribution ahead of the descending thaw front is supported by observations of summer heave in the frozen active layer by Mackay (1983).

Fall freeze-back of the active layer involves penetration of the zero degree isotherms from the surface downward and from the permafrost upward (Mackay, 1983). Moisture is drawn by thermally-induced suction gradients from the central portion of the active layer towards the descending and ascending thaw fronts resulting in the growth of ice lenses subparallel to the ground surface and at the base of the frozen active layer (Mackay, 1983). Ice-lens development during freeze-back is concentrated principally in the upper decimeters of the active layer (Cheng, 1983). At the end of freeze-back, the central portion of the active layer is frequently desiccated from water loss to the two freezing planes (Parmuzina, 1978; Mackay, 1983). The magnitude and duration of winter frost heave is constrained by the

Figure 2.5 Soil water content change in fine-grained soils over summer 1976 at Fenghuo Shan, P.R.C. (Cheng, 1983, Figure 7). (1) Soil water content, May 22; (2) October 2; and (3) maximum depth of seasonal thawing.

availability of moisture within the active layer. Intrusive ice may result from the freezing of water that has been injected into the active layer by cryostatic pressure in an adjacent water body (e.g. ice-wedge trough) during the autumn freeze-back period (Mackay, 1972, 2000). Injection ice may be present as discrete masses of pure ice at the base of the active layer or may become incorporated into permafrost (Mackay, 2000).

In winter, once freeze-back is complete, the thermal gradient in the active layer and near-surface permafrost is reversed and temperatures decrease upwards. Sublimation losses from the frozen soil may desiccate the upper part of the active layer producing a dry friable surface soil layer conducive to infiltration of spring meltwater (Mackay, 1983). Measurements of vapour flow out of frozen soil during winter ranged from between 3 mm (Marsh and Woo, 1984; Smith and Burn, 1987) to about 30 mm (Santeford, 1978).

During winter, slow heave within the active layer continues due to the freezing of pore water and upward redistribution of moisture along the thermally-driven suction gradient (Mackay, 1983). Vapour diffusion through pore spaces in the desiccated portion of the active layer (Hinkel and Outcalt, 1994) may also contribute to heave at the top of the frozen active layer (Mackay, 1983). Parmuzina (1978) reported moisture loss from permafrost and noted desiccation of the upper 10 cm of permafrost during late winter.

2.8 Aggradational ice

The development of the near-surface ice-rich zone is linked to the thermal regime of permafrost and processes of moisture migration in freezing and frozen soils (Cheng, 1983; Mackay, 1983). This segregated near-surface ground ice was termed aggradational ice by Mackay (1972). Shur (1988) described the ice-rich zone as containing a volumetric ice content generally over 50% with abundant ice lenses. The higher ice content and thicker lenses in near-surface permafrost provide a distinguishing boundary between permafrost and the overlying active layer (Mackay, 1972).

There are two main hypotheses that explain the presence of aggradational ice. Mackay (1972) attributed its origin to the trapping of ice lenses at the base of the former active layer by a rising permafrost table. A rise in the permafrost table could occur as a result of climate change, vegetation succession or sedimentation. Cheng (1983) suggested that this ice-rich zone is a result of the seasonal water balance of permafrost terrain. Water may move downward along a temperature-induced suction gradient at the end of summer when the permafrost is warm and soil hydraulic conductivity is relatively high (Horiguchi and Miller, 1983). The annual downward flux may be greater than water movement out of this zone in winter (Parmuzina, 1978), when the temperature gradient is reversed (Cheng, 1983). The net imbalance between summer and winter vertical moisture fluxes is attributed to seasonal changes in unfrozen water content and hydraulic conductivity of frozen soil (Figure 2.6) (Cheng, 1983). Mackay (1983) supports the notion that the

Figure 2.6 The relationship between ground-temperature envelopes and the curve of unfrozen moisture content vs. soil temperature. A) Region of positive ground-temperature gradient; B) region of negative ground-temperature gradient (Cheng, 1983, Figure 4).

direction and magnitude of moisture migration has seasonal components, but that over a short duration, the influx of moisture into the upper permafrost should be nearly balanced by the efflux.

Downward migration of water into permafrost is supported by seasonal observations of moisture migration and frost heave (Parmuzina, 1978; Cheng, 1983; Mackay, 1983; Smith, 1985; Burn, 1988; Hinkel *et al.*, 1996; Mackay, 1997). Mackay (1997) measured near-surface heave in a recently drained lake bottom, which he attributed to the formation of aggradational ice as the permafrost table aggraded in response to vegetation succession. Burn (1986) measured an annual net increase in water content at the top of permafrost in the order of a few mm a^{-1} in process-related studies conducted near Mayo, YT. Using soil analogue tubes, Burn (1988) measured significant amounts of moisture migration over a single season (10 mm a^{-1}) into the upper 20 cm of perennially frozen ground. However, Mackay has indicated that the initial rates of moisture movement into frozen ground may decrease over time (pers. comm. in Burn, 1988).

Hinkel *et al.* (1996) investigated moisture accumulation in near-surface permafrost over 30 years at a site near Barrow, Alaska. The study involved statistical comparisons of moisture content in near-surface permafrost between cores obtained in 1963 and paired cores collected in 1993. A statistically significant increase of 5% in volumetric ice content over thirty years was determined. Analysis of annual change in moisture content of the upper permafrost using the same techniques was inconclusive.

The migration of water into permafrost has been inferred using tritium (^3H) as a tracer (Michel and Fritz, 1982; Chizhov *et al.*, 1983; Burn and Michel, 1988).

Tritium originates from nuclear testing in the 1950's and 1960's, therefore it has been assumed that the depth of moisture migration since 1950 is given by the tritium profile. Elevated levels of tritium 10 to 30 cm below the permafrost table have been attributed to the downward migration of atmospherically-derived moisture into permafrost (Chizhov *et al.*, 1983; Burn and Michel, 1988). Relatively deep infiltration of tritium into fine-grained sediments reported by Michel and Fritz (1978) and Burn and Michel (1988) at sites that were irrigated with tritium are consistent with the higher unfrozen water content and hydraulic conductivity of these materials compared to coarse-grained soils at temperatures below 0°C.

Mackay's (1972) hypothesis of aggradational ice development is supported by field evidence (e.g., Burn, 1988), but it implies that the ice-rich zone can only develop under conditions of aggrading permafrost. Cheng's (1983) hypothesis, based on water migration through frozen ground, suggests that the upper part of permafrost is a net sink for active-layer water. Cheng's hypothesis implies that moisture will continue to migrate through growing ice lenses (Figure 2.3), however, there is little evidence indicating that this process can be sustained as the upper zone of permafrost becomes increasingly ice rich.

2.9 Solute redistribution during ground freezing

During the freezing of saturated soils, solute rejection from the ice phase will redistribute solutes in unfrozen soil ahead of the advancing freezing front (Hallet, 1978; Mackay, 1998). At a micro-scale, solute exclusion during soil freezing will enrich soluble ion concentrations on the warm side of the ice/water interface. If the temperature is lowered, ice nucleation may occur at some distance away from the freezing interface where solute concentrations are lower and the freezing-point depression is less (Hallet, 1978). The degree of solute exclusion from the ice phase is a function of the freezing rate (Hallet, 1978).

In laboratory experiments with sandy soil, Baker and Osterkamp (1988) measured solute enrichment in unfrozen soil beneath a descending frost front. The absence of an interfacial zone of solute enrichment indicated convection in the underlying thawed region and that the convective velocity of the salt was greater than the maximum freezing rate in the soil (Baker and Osterkamp, 1988). Kay and Groenevelt (1983) and Baker and Osterkamp (1988) found that there was no effective salt exclusion from the freezing soil during upward freezing. These results indicate that gravity-driven brine drainage in the frozen fringe acts to dissipate interfacial solute concentrations, improving the efficacy of solute exclusion during downward freezing.

Several laboratory experiments conducted with fine-grained silt and clay demonstrated that there is net movement of moisture and ions towards the freezing zone (Figure 2.7) (Cary and Mayland, 1972; Qui *et al.*, 1988; Chuvilin *et al.*, 1998),

Figure 2.7 The redistribution of soluble salts during the freezing of silty clay (Qui *et al.*, 1988, Figure 3).

however field evidence of this phenomenon is limited (Kay and Groenevelt, 1983).

Qui *et al.* (1988) concluded that in most silts and clays, ion redistribution due to moisture migration and temperature diffusion towards the freezing front would greatly exceed the amount of concentration diffusion away from the frozen zone.

The exclusion of solutes from water drawn towards a growing ice lens results in a zone of ionic enrichment on the warm side of a growing lens (Hallet, 1978). In tests on the freezing of sands containing fines, solute exclusion resulted in unbonded brine-rich zones between ice lenses (Chamberlain, 1983). High interfacial solute concentrations may be expected during the growth of aggradational ice lenses because of the absence of gravity-driven brine drainage during upward freezing (Baker and Osterkamp, 1988). Therefore, aggradational ice and intervening soil bands may be enriched in soluble ions.

Data presented by O'Sullivan (1966) and Ostroumov *et al.* (1998) suggested that the zones of ionic and moisture enrichment in the upper part of permafrost generally coincide. Tarnocai (1972) attributed elevated Ca^{++} concentrations in frozen peat to downward moisture and nutrient migration along the thermal gradient.

Coupled movement of moisture and solutes during the freezing of fine-grained soils and the absence of solute exclusion during upward freezing suggests that the near-surface ice-rich zone could be enriched in solutes relative to mineral soils at the base of the active layer. The accumulation of soluble ions in the ice-rich zone is of ecological significance because surface disturbance or climate warming could result in thawing of permafrost and the release of nutrients previously held in the

near-surface permafrost (Leibman and Streletskaya, 1997; Kokelj and Lewkowicz, 1999; Kokelj *et al.*, 2002).

CHAPTER 3. REGIONAL SETTING

This chapter describes the physical environment of the Mackenzie Delta and the upland region adjacent to Inuvik and provides a context for the research and experimental techniques.

This study was conducted on the eastern side of the Mackenzie Delta and on terrain adjacent to the Delta near Inuvik (Figure 3.1). The Mackenzie Delta is a Holocene feature that is bounded by the Richardson Mountains to the west and the Caribou Hills to the east. The Delta is approximately 150 km long, up to 60 km wide and is comprised mainly of alluvial silts deposited by the Mackenzie and Peel rivers. The elevated bench along the east shore of East Channel near Inuvik is a low-angled fan deposit with materials derived from hills and gullies to the east (Heginbottom and Tarnocai, 1983). Surficial materials on this surface are silt and clay with coarse gravels, but colluvial and morainal deposits are also present (Rampton, 1988).

3.1 Regional Quaternary history

Stratigraphic and geomorphic records preserved in the Richardson and Mackenzie mountains suggest that the Wisconsinan ice sheet was the only continental glaciation to reach these areas (Duk-Rodkin and Lemmen, 2000). Prior to the Wisconsinan glaciation, water in the Mackenzie Basin flowed east, contributing to the Bell River paleodrainage system which flowed through present-day Hudson Bay and into the North Atlantic (Duk-Rodkin and Hughes, 1994). At

Figure 3.1. Map of the Mackenzie Delta region. The study areas in southern, central and northern (outer) Mackenzie Delta are indicated on the map: SD – southern Mackenzie Delta; CD - central Mackenzie Delta; OD - outer Mackenzie Delta. N indicates the Navy Road study area on terrain adjacent to the Delta near Inuvik. D1 is a spruce/crowberry-lichen study site referred to in Chapter 7.

that time, only the Peel and Anderson river basins drained into Beaufort Sea (Duk-Rodkin and Lemmen, 2000). The Wisconsinan glaciation and subsequent erosion resulted in a massive diversion of drainage and formation of the modern Mackenzie Basin (Duk-Rodkin and Hughes, 1994).

Radiocarbon dates from sub-till samples in northwest Canada suggest that in this area, the Laurentide maximum occurred as early as 30 ka BP (Dallimore *et al.*, 1997). During this period, glacial ice covered the area now occupied by Mackenzie Delta, Richards Island and much of the Tuktoyaktuk Coastlands (Figure 3.2) (Rampton, 1988; Dallimore *et al.*, 1997). Ice thrusting resulted in the excavation of Herschel Island along the Yukon Coastal Plain, indicating the maximum northwest extent of this ice sheet (Rampton, 1982). At the glacial maximum, Wisconsinan ice bordered on the Mackenzie and Richardson Mountains, creating a complex system of lakes and rivers that drained northwestwards. This diversion in drainage led to the inundation of the Bell, Bluefish and Porcupine basins (Duk-Rodkin and Lemmen, 2000).

By 20 ka BP, Tuktoyaktuk Peninsula, Richards Island and part of the Beaufort Shelf were free of Laurentide ice (Dallimore *et al.*, 1997). The continental ice mass and montane glaciers had retreated from the Richardson Mountains by 13 ka BP (Tutsieta Lake phase) but ice still occupied the Mackenzie Valley and the Delta trench (Duk-Rodkin and Lemmen, 2000). By 12.5 ka BP, the ice sheet had retreated and proglacial Lake Travaillant formed in the Mackenzie Valley just southeast of Tsiigehtchic. The Mackenzie corridor was free of ice by 10 ka BP, but the region

Figure 3.2 Ice flow and glacial limits during the Early Wisconsin Toker Point stage (Buckland Glaciation) (Rampton, 1988, Figure 55).

was dominated by Glacial Lake McConnell, which occupied all of the modern day Great Bear, Great Slave and Athabasca lake basins (Duk-Rodkin and Lemmen, 2000).

3.2 Local Quaternary history and Holocene paleoclimate

Glacial fluting and striations indicate that the trough occupied by the modern Mackenzie Delta was a major flow channel of a lobe of the Laurentide ice sheet (Rampton, 1972). West of the Delta, the glacial limit decreased from an elevation of 90 m in the Richardson Mountains near Aklavik to present sea-level near Herschel Island (Rampton, 1972). The Toker Point Stade ice advance covered the entire Mackenzie Delta, Richards Island and much of the Tuktoyaktuk Coastlands sometime after 33.7 ka BP (Figure 3.2) (Rampton, 1988; Dallimore *et al.*, 1997). The ice cover probably resulted in some degradation of subglacial permafrost (Rampton, 1988). There is, however, some discrepancy as to the extent of ice north of Tuktoyaktuk, as till and erratics from this region have been obscured by outwash and marine transgression (Rampton, 1988). It is believed that the coastal islands to the north of the Mackenzie Delta are ice thrust features as the result of this glaciation (Mackay and Stager, 1966). Mackay (1956) deduced that permafrost was already present in these sediments at the time of the glacial advance, as both sediments and ground ice are deformed.

Following the deglaciation of Richards Island, it is believed that Laurentide ice occupied the Mackenzie Delta trench for a significant period during which time

elevated terraces, such as those near Tununuk, were formed by meltwater flowing down East Channel (Rampton, 1988). At this time, the crests of the Caribou Hills were also free of Wisconsinan ice.

Upstream of the Delta, the Mackenzie River flows through sediments deposited by proglacial lakes that occupied the corridor between 14 and 10 ka BP (Duk-Rodkin and Lemmen, 2000). Immediately following glaciation, meltwater from glacial Lake Travaillant drained through the Anderson River Valley, then into the Delta via the valley presently occupied by the Rengleng River. By about 14 ka BP, water had cut a canyon through bedrock in the vicinity of Tsiigehtchic, and melt water from glacial lakes up the valley began to flow into the Delta (Duk-Rodkin and Lemmen, 2000).

The Mackenzie Delta evolved over the last 13,000 years (Hill *et al.*, 1996). As continental ice melted and sea level rose, sediment transported by glacial meltwater began building the Delta in a seaward direction. Much of the sediment deposited in the Delta is derived from lacustrine sediments in the Mackenzie and Peel river valleys. Several small valleys draining the Caribou Hills were incised when sea levels were approximately 100 m lower than at present (Rampton, 1988). These “drowned valleys” infilled with sediment as sea levels rose and the Delta prograded (Rampton, 1988). At present, the Delta is in a transgressive stage (Hill *et al.*, 1996).

Paleoclimatic investigations in the western Arctic indicate that the early Holocene climate was warmer than it is today (Ritchie, 1984; Rampton, 1988). Warmer temperatures were due to increased summer insolation attributed to

Milankovitch cycles, and also to change in the paleogeographic configuration of the coastline along the continental shelf of the Beaufort Sea (Ritchie *et al.*, 1983; Burn, 1997).

Paleoecological studies indicate a rapid climate warming between 12,000 and 10,000 BP (Ritchie, 1984). During this period, mean daily temperatures in the summer months may have been 3-5°C higher than at present (Ritchie, 1984). By 9500 BP, treeline had advanced onto the Tuktoyaktuk Peninsula, and was at least 100 km north of its present location where it remained until its retreat after 4900 BP when climate deteriorated (Ritchie and Hare, 1971).

The warmer climate led to an increase in thaw depth and formation of a widespread thaw unconformity (Mackay, 1975; Burn, 1997). In the Mackenzie Delta area, a considerable proportion of the increase in summer thaw depth was likely due to northward extension of the taiga, leading to deeper snow cover and warmer ground temperatures (Burn, 1997). During the height of the postglacial warm period, extensive thermokarst activity took place in the Mackenzie Delta region (Rampton, 1988; Burn, 1997; Murton, 2001). The active layer deepened regionally, with maximum thickness inferred from ^{14}C estimates of *in situ* rootlets above the base of the thaw unconformity occurring around 8000 BP (Burn, 1997). In the western Arctic, truncated ice wedges, variations in the cryostructure and ice content of ground ice, and a geochemical discontinuity are indicative of the thaw unconformity (Mackay, 1975; Murton and French, 1993; Kokelj *et al.*, 2002).

At present, there is a steep environmental gradient in growing season conditions between sites near the Beaufort Sea and warmer inland locations (Burn, 1997). The paleoactive layers identified along the present day coast in the Mackenzie Delta region developed at sites that were about 100 km inland during the early Holocene. Since that time, sea level north of the present Mackenzie Delta has risen about 70 m (Hill *et al.*, 1993) and one- to two-thirds of summer climate cooling experienced after 8000 BP along the present-day coastline may be attributed to coastal recession (Burn, 1997).

A more recent climatic amelioration known as the “Medieval Warm Period” (Hughes and Diaz, 1994) is indicated by tree-ring evidence from the Eskimo Lakes area (Figure 3.1). The tree-ring chronology indicates favourable growing conditions from 1172 to 1330 BP (Bégin *et al.*, 2000). The tree growth suggests that summers were warmer than today and the shape of subfossil trees indicate that winters were also milder with greater snowfall than at present (Bégin *et al.*, 2000).

3.3 Contemporary climate

The Mackenzie basin has a continental climate with long cold winters and relatively warm summers. Mean monthly temperatures for a north-south transect in the Mackenzie Valley indicate that monthly means between Inuvik and Fort Good Hope vary by less than 5°C (Dyke, 2000). Cold winter temperatures throughout the lower Mackenzie Valley are due to an extensive high-pressure ridge causing stable weather patterns to persist (Table 3.1a). Climate in the Delta is transitional between

a)	Temperature	J	F	M	A	M	J	J	A	S	O	N	D	Year
Fort McPherson	Mean (°C)	-27.2	-24.3	-20.2	-10.2	1.9	12.9	15.5	11.7	4.5	-7.7	-21.4	-25.1	-7.5
Inuvik	Mean (°C)	-28.3	-27.6	-23.7	-13.2	-0.3	10.9	13.9	10.9	3.5	-8.1	-21.1	-26	-9.1
Tuktoyaktuk	Mean (°C)	-26.9	-27	-25.8	-16.7	-4.9	5.9	11	9	2.9	-8.1	-20.5	-25.2	-10.5

b)	Precipitation	J	F	M	A	M	J	J	A	S	O	N	D	Year
Fort McPherson	Total (mm)	16	13	12	11	17	26	46	38	37	27	27	15	285
	Snow (cm)	16	13	12	11	11	Tr	0	Tr	8	37	27	15	150
Inuvik	Total (mm)	16	12	12	12	17	21	34	42	25	30	17	16	254
	Snow (cm)	18	15	15	15	13	2	Tr	3	11	34	22	20	168
Tuktoyaktuk	Total (mm)	10	10	6	9	7	10	21	29	24	20	12	9	167
	Snow (cm)	10	10	6	9	6	2	Tr	Tr	8	20	12	9	92

Table 3.1 a) Mean monthly temperatures and; b) total precipitation and snowfall at Fort McPherson (1981-2001), Inuvik (1957-2001) and Tuktoyaktuk (1970-2001) (Meteorological Service of Canada, 2002).

arctic conditions in coastal areas and subarctic conditions to the south. The region between Reindeer Station and Tununuk is in the transitional zone (Figure 3.1) (Mackay, 1963). The Beaufort Sea has a significant effect on spring and early summer temperatures in the Delta region. Persistent ice cover retards the onset of spring and shortens the length of the growing season at sites within 50 km of the coast (Figure 3.3) (Burn, 1997). For example, mean June temperature is 12.9° at Fort McPherson, 10.9°C at Inuvik, but only 5.9°C at Tuktoyaktuk (Table 3.1a). Temperature differences between Tuktoyaktuk and Inuvik diminish later in the summer once the sea becomes ice-free (Table 3.1a).

Precipitation in the Mackenzie Valley is generally low due to the rainshadow effect east of the Cordillera. Total annual precipitation varies from about 300 to 400 mm in the Mackenzie Valley, but decreases to 254 mm in the Delta at Inuvik and 167 mm at Tuktoyaktuk (Table 3.1b). From late October to April, precipitation in the Mackenzie Delta region falls as snow. Mean snowfall in Inuvik is about 168 cm while Tuktoyaktuk receives less than 100 cm (Table 3.1b). The lower amount of snowfall at the coast is attributed to the dominance of dry arctic air and the greater distance from moisture-carrying storms which track further to the south (Dyke, 2000). In the Delta region, the heaviest snowfalls occur in fall (Table 3.1b) prior to the establishment of the arctic high-pressure ridge that occupies the Valley later in winter (Dyke, 2000).

Figure 3.3 Mean annual total growing degree-days and distance from the Arctic coast for stations in Yukon Territory and western Northwest Territories. Growing degree-days ($>5^{\circ}\text{C}$) calculated from mean monthly air temperatures in Environment Canada (1982) (Burn, 1997, Figure 10).

3.4 The Mackenzie Delta

3.4.1 Fluvial hydrology

The Mackenzie Delta is characterized by an intricate network of channels (distributary, river, network, reversing and tidal) and lakes (abandoned, arcuate, floodplain, thermokarst, dammed) of different type and origin that comprise between 15 and 50% of its total surface area (Mackay, 1963). The Mackenzie and Peel rivers supply the majority of sediments to the Delta, the bulk of which are derived from Laurentide glaciolacustrine, morainal and till deposits.

Annually, the Mackenzie River contributes about 100 million tonnes of sediment to the Delta and the Peel River supplies between 15 and 25 million tonnes (Carson *et al.*, 1998). Mean annual net sedimentation in the Delta is estimated by Carson *et al.*, (1999) at about 40 million tonnes and outflow at about 80 million tonnes. Borehole and seismic data have provided estimates of sedimentation rates during the Holocene of between 136 to 163 million tonnes per year (Hill *et al.*, 1991). Some of the outflow estimated by Carson *et al.* (1999) would be deposited as foreset beds at the Delta mouth, thus totals including sedimentation and outflow are roughly comparable with the estimates made by Hill *et al.* (1991).

The Mackenzie River contributes about 90% of the total discharge into the Delta, with the balance supplied by the Peel River. Mean discharge of the Mackenzie is $9140 \text{ m}^3 \text{ s}^{-1}$ (Brooks, 2000). Winter discharge ranges between 4000 and $5000 \text{ m}^3 \text{ s}^{-1}$, during which time much of the water in the basin is stored as snow and ice (Brooks, 2000). The peak in Mackenzie River discharge usually occurs in

May or early June. The rapid influx of water from the Liard River in spring induces breakup of ice in the upper Mackenzie system (MacKay and Mackay, 1973). As southerly tributaries reach peak spring discharge, ice in the Delta may still be intact. The flood wave from the south eventually reaches the Delta and exerts significant pressure on the ice cover, resulting in the lifting and fracturing of the ice mass (MacKay and Mackay, 1973). Ice jams severely obstruct river flow and can result in a rapid rise in upstream water levels of up to 7 m above winter ice levels in East Channel and connected lakes of the southern Delta (Bigras, 1990). More commonly, the flood levels range between 3 and 6 m above winter ice levels (Bigras, 1990). Levee heights in the Delta decrease northwards, consequently, a greater proportion of the Delta surface is inundated by spring flooding as distance below Point Separation increases (Marsh and Hey, 1991).

The thermal effect of flooding is to eradicate ice cover from channels and lakes in the Delta. Flooding removes snow cover, warming the ground and accelerating active-layer development. Floodwaters rich in Ca^{++} and Mg^{++} (Lesack *et al.*, 1991) and alluvium deposited during floods contribute nutrients to the terrestrial system enhancing productivity of Delta ecosystems with respect to those on the adjacent tundra.

The Mackenzie Delta is a lake-rich environment with over 25 000 lakes, which have been classified by the frequency with which they are flooded (Mackay, 1963). No-closure lakes are connected to the main channel for the entire year (Marsh and Hey, 1991; Burn, 1995). Low-closure lakes are flooded each spring but cut off from

the river for some portion of the year. High-closure lakes are not flooded every spring and are defined by sill elevations greater than the one-year return period for spring peak water levels. The percentage of lakes that flood annually increases downstream.

Hydrological and sedimentary regimes vary between lakes with similar sill elevations (Ferguson and Marsh, 1991). For example, many of the lakes are not connected directly to adjacent channels, but rather by a series of lakes and creeks. Since flow directions in these creeks may vary throughout the summer based on fluctuation in channel stage, lakes distant from channels are influenced by different sedimentary regimes than those with close hydrological connections to the channel. Storm surges can produce short-term increases in water levels and intrusion of salt water into channels and lakes of the outer Delta (Jenner and Hill, 1991). Flow reversals result in the deposition of sediment, accretion of emergent bars and infilling of small channels (Jenner and Hill, 1991).

3.4.2 Permafrost and geomorphology

The Mackenzie Delta consists of alluvial sediments that have accumulated over the past 13 000 years (Hill *et al.*, 1996). A deep bore hole drilled southeast of Inuvik penetrated 54 m of alluvial sands and silts, 8 m of dense clay and 7 m of pebbly clay before hitting bedrock (Johnston and Brown, 1965). At least 150 m of unconsolidated sediments were penetrated in a drill hole on Ellice Island in the outer Delta (Rampton, 1972).

Most terrestrial surfaces in the Delta are undergoing active sedimentation or erosion (Mackay, 1963; Gill, 1972). Terrestrial sedimentation rates vary with frequency and duration of flooding (Marsh and Hey, 1991). Pearce (1993, 1994) measured sedimentation rates on various Delta surfaces and characterized sedimentation by vegetation community. Annual sedimentation rates were measured from 1981 to 1983 and in 1992 and 1993. At the point-bar study site in central Mackenzie Delta (CD) (Figure 3.1), annual accumulation of sediment on mudflats adjacent to the stream channel was 20 cm a^{-1} (Pearce, 1993, 1994). Sedimentation on aggrading point bars colonized by pioneer willow communities ranged from 5 to 20 cm a^{-1} . Lower rates of sedimentation, between 0 and 10 cm a^{-1} were measured at mature willow stands that colonize higher elevations. In years that floodwaters inundate spruce forests of the Delta plain, sediment accumulation on these elevated surfaces, is usually only a few mm (Mackay, 1963; Pearce *et al.*, 1988; Pearce, 1993, 1994). Annual sedimentation on a lakeshore colonized by alder was between 0 and 0.5 cm a^{-1} (Pearce, 1993, 1994). In the vicinity of the outer Delta study area (OD) (Figure 3.1), sedimentation rates on aggrading point bars colonized by willows were between 2 to 10 cm a^{-1} , but at mesic sites colonized by sedges, distal from major channels, sedimentation was less than 2 cm a^{-1} (Pearce, 1993, 1994).

Permafrost in the Mackenzie Delta is discontinuous and is generally less than 100 m thick due to the thermal effect of water bodies and the shifting nature of the channels (Smith, 1976). As channels migrate or lakes expand due to thermokarst

processes, so too do the underlying taliks (Figure 3.4). Channel shifting ultimately results in the emergence of the river bottom, which forms a subaerial bar where permafrost is eventually re-established. Permafrost thickness increases with distance away from the migrating channel, as the degree of thermal recovery increases (Figure 3.4) (Smith, 1975). Factors such as vegetation succession and snow accumulation influence the microclimate of point bars, which affects the ground thermal conditions and active-layer thickness (Smith, 1975).

Fine-grained sediments and a rising permafrost table encountered in an alluvial environment are conducive to the development of near-surface aggradational ice (Mackay, 1972), but the spatial variability in near-surface ground ice in Mackenzie Delta is not known. Little or no segregated ice was observed in Delta sediments obtained from a drill core that extended to a depth of 70 m (Johnston and Brown, 1965; Williams, 1968). Investigations near Aklavik indicate that segregated ice is concentrated in the upper 2 to 3 m of alluvial sediments (Pihlainen and Johnston, 1954). Permafrost deeper in the core profiles was ice-poor with little or no excess ice (Pihlainen and Johnston, 1954).

Ice-wedge polygons are present on elevated surfaces throughout the Delta (Mackay, 1963). Sedimentation and aggradation of the permafrost table cause these wedges to grow upwards (Mackay, 1990), but deposition of alluvium and vegetative change obscure the polygons, so troughs are difficult to detect. Some small pingos have formed where permafrost has aggraded into sediments of infilled lake-basins or abandoned channels (Mackay, 1963). The pingos of the outer Mackenzie Delta

Figure 3.4 Diagrammatic cross-section of a shifting channel showing the configuration of permafrost and point-bar vegetation types, central and southern Mackenzie Delta (modified after Smith, 1975, Figure 4).

are generally much smaller than those of the Tuktoyaktuk Coastlands, but remain conspicuous due to the low surrounding relief.

3.4.3 Vegetation communities in the Mackenzie Delta

Channel migration results in a complex interaction between topography, vegetation and microclimate and thermal regime (Gill, 1972; Smith, 1975). Vegetation succession on aggrading point bars is initiated by a horsetail (*Equisetum*) association (Gill, 1973). A lens of permafrost is often present because the removal of snow by winter winds exposes ground to cold winter temperatures (Figure 3.4). Willow-horsetail (*Salix-Equisetum*) associations are encountered with increasing distance from the stream channel. The soils are sandy silts and active layers are between 100 and 150 cm thick. Permafrost is sometimes absent because dense willows trap snow which retards winter heat loss from the ground (Smith, 1975). In point-bar environments of the central and southern Delta, alder (*Alnus*) colonizes wet micro-environments such as meander scroll depressions (Gill, 1973). An active layer 90-120 cm thick forms in the silty alluvium.

In the central and southern Delta, poplar (*Populus balsamifera*) and white spruce (*Picea glauca*) colonize elevated surfaces (Mackay, 1963). Spruce succession is associated with a reduction in flooding and sedimentation, which permits establishment of a moss layer, while the development of a canopy reduces snow cover (Gill, 1975). These changes contribute to colder ground temperatures and thinner active-layer depths (Smith, 1975).

Pearce *et al.* (1988) identified three stages of spruce succession in the Mackenzie Delta. Initial colonization by spruce occurs on land surfaces 3 to 4 m above low water level in the middle Delta, and 7 to 8 m above low water in the southern Delta. Trees are straight and grow relatively rapidly in association with an alder and willow shrub layer. Flooding is frequent, but white spruce survive sedimentation and surface aggradation by developing adventitious roots (Gill, 1975). The soils are static cryosols and the permafrost table is encountered between 80-100 cm depth. As the elevation of the land surface increases, flood frequency declines and the spruce/alder-bearberry (*Picea/Alnus-Arctostaphylos*) stands are replaced with spruce/feathermoss (*Picea/Hylocomium*) communities. Individual trees are commonly older than 250 years and have survived centuries of flooding, sedimentation and aggradation of the permafrost table. Consequently, their trunk and root systems are anchored in permafrost (Gill, 1975). Active-layer depths are between 50 and 70 cm. Spruce/crowberry-lichen (*Picea/Empetrum-Cladina*) communities occupy land surfaces above the level of seasonal flooding between 4.5 to 6 m above low water in the Inuvik region and on surfaces greater than 8 m above low water in the southern Delta. The spruce trees are stunted and lean in random directions. The soils are organic cryosols with active-layer depths between 20 to 60 cm.

Lakeshore environments tend to flood frequently, but mean annual sedimentation rates are about an order of magnitude lower than on aggrading point bars (Pearce, 1993, 1994). These poorly drained sites are colonized by alder in the

southern and central Delta, and by sedge tussocks in the outer Delta. Active-layer depths vary from about 40 to 90 cm.

3.5 Inuvik area adjacent to the Mackenzie Delta

A narrow bench runs between the Mackenzie Delta and the Caribou Hills. The Navy Road study site, approximately 2 km north of Inuvik is located in this terrain (Figure 3.1). Surficial materials are low-angle fan deposits composed predominantly of clayey silts derived from hills and gullies to the east (Rampton, 1988). Gravels, interpreted as kame-terrace deposits, were laid down between the hills and a late glacial ice lobe that occupied the present position of the Delta (Rampton, 1988).

Permafrost in the vicinity of Inuvik is approximately 100 m deep (Burn and Kutny, pers. comm., 2002). Ice wedges are present, but earth hummocks are the most common soil micro-relief feature in the area (Mackay, 1995). Soil types are variable, but most are frost-susceptible and contain high ice contents (Heginbottom, 1972; Mackay, 1995). Estimated excess ice contents in near-surface permafrost range from 40 to 90% (Heginbottom, 1974; Watson *et al.*, 1973). Active-layer depths vary between 30 to 150 cm beneath earth hummocks where the permafrost table is bowl-shaped (Mackay, 1995).

Vegetation in the region varies with moisture availability and disturbance history. An understorey of *Rhubus chamaemorus*, *Petasites frigidus*, *Ledum decumbens*, *Vaccinium vitis-idaea* and an associated thick moss cover is

characteristic of undisturbed sites near Inuvik (Mackay, 1995). Black spruce (*Picea mariana*), white spruce (*Picea glauca*) and paper birch (*Betula papyrifera*) are also present. Fireweed (*Epilobium angustifolium*) and other disturbance species are the first to colonize disturbed sites. Thirty years after a fire, Mackay (1995) observed that willow (*Salix*), alder (*Alnus*) and paper birch had re-colonized the affected area, but that spruce was absent.

CHAPTER 4. RESEARCH DESIGN AND METHODS

This chapter summarizes the research design with reference to the specific study sites and describes the field and laboratory methods implemented to investigate the physical and chemical characteristics of near-surface permafrost in the Mackenzie Delta region.

4.1 Research design

The objectives of this thesis were investigated using the following research techniques:

Objective 1. To examine the spatial distribution of near-surface ground ice in sediments of the Mackenzie Delta.

The near-surface ice-rich zone is widespread, but not ubiquitous in permafrost terrain (Mackay, 1972; Burn, 1988). The spatial distribution of this ice-rich zone in sediments of the Mackenzie Delta is not well known. Soil properties such as grain size affect the frost susceptibility of freezing and frozen soils and the development of near-surface segregated ice lenses (Williams, 1968), but other factors such as soil-moisture conditions, rates of permafrost aggradation (Mackay, 1972) and the ground-thermal regime (Cheng, 1983) may also influence development of ice-rich permafrost.

It was hypothesized that in sediments of Mackenzie Delta, sediment texture controls segregated ice accumulation in near-surface permafrost. To test this hypothesis and to identify other physical characteristics that favour aggradational

ice development transects were established perpendicular to shifting channels in the southern Mackenzie Delta (Figure 4.1), in the mid-Delta near Reindeer Station at Gill's camp (Figure 4.2), and in the outer Delta (Figure 4.3) (also see Figure 3.1 for general location). A total of 43 shallow permafrost cores were analyzed to assess near-surface ground-ice conditions and soil physical properties along the environmental gradients associated with point-bar environments (Gill, 1972; Smith, 1975; Pearce, 1993, 1994). On point bars in southern and central Mackenzie Delta, permafrost samples were obtained from willow, alder, poplar and spruce communities (Figures 4.1 and 4.2). In the outer Delta, transects traversed point-bar willow communities. Active-layer and near-surface ground-ice conditions were also determined in other terrain types encountered in the Delta. For example, permafrost cores were obtained from lakeside alder and sedge communities in the central and outer Delta, respectively (Figure 4.2 and 4.3).

Pervious work by Pearce (1993, 1994) on sedimentation rates and Pearce *et al.*, (1988) on vegetation succession at the southern (SD), central (CD) and outer Delta (OD) study sites, and by Gill (1972, 1973, 1975) on vegetation succession, and Smith (1973, 1975, 1976) on ground temperatures and microclimate at the study site in central Mackenzie Delta, provide a context within which to discuss near-surface ground-ice conditions determined by coring. Since the conditions encountered at the respective study sites represent the range of soil, moisture, flooding and ground-thermal environments encountered in the Delta (Gill, 1972; Smith, 1975; Pearce, 1993, 1994), the results presented in this thesis are applicable throughout the

Figure 4.1 Map of the southern Mackenzie Delta study area (SD). See Figure 3.1 for general location ($67^{\circ}52'N$, $134^{\circ}09'W$). Core locations are indicated. Core S-w1 is from a willow community; S-a1 is from an alder community; S-p1 is from a poplar community; S-sab1 is from a spruce/alder-bearberry community; and S-sf1 and S-sf2 are from a spruce/feathermoss forest. Core code indicates (S-sab1): a) General location, **S** – southern Delta; b) forest type, **sab** – spruce/alder-bearberry forest; and c) core number, **1**.

Figure 4.2 Map of the central Mackenzie Delta study area, Gill's Camp (CD). See Figure 3.1 for general location ($68^{\circ}43'N$, $134^{\circ}16'W$). Core locations are indicated along the transect. Cores C-w1 to C-w6 are from point-bar willow communities; C-a1 and C-a2 are from alder stands; C-sab1 and C-sab2 are from a spruce/alder-bearberry forest; C-sf 1 to C-sf7 are from a spruce/feathermoss forest; and C-la1 to C-la3 are from a lakeshore alder community.

Figure 4.3 Map of the outer (northern) Mackenzie Delta study area (OD). See Figure 3.1 for general location ($69^{\circ}10'N$, $135^{\circ}27'W$). Core locations are indicated by an X. Cores O-w1 to O-w5 are from point-bar willow communities; O-s1 and O-s2 are from sedge communities; and O-Pin1 and O-Pin2 are from a small pingo.

Mackenzie Delta and to a range of other permafrost environments.

Objective 2. To investigate vertical patterns of near-surface ground-ice accumulation with respect to aggrading, degrading and static permafrost tables.

Cheng's (1983) hypothesis suggests that under equilibrium conditions, near-surface permafrost will become enriched with ice. However, the hypothesis does not define the vertical extent of this ice accumulation, and whether this process is sustained as the ice-content of permafrost increases. Mackay (1972) suggested that a rising permafrost table may trap ice lenses formerly at the base of the active layer. Therefore, an aggrading permafrost table associated with sedimentation, climate change or vegetation succession may result in the development of a thick ice-rich zone (Mackay, 1972; Cheng, 1983).

It was hypothesized that under static permafrost table conditions, ice accumulation is limited to the top 10 cm of permafrost. Vertical distribution of near-surface segregated ice was investigated at a site in hummocky terrain approximately 2 km north of Inuvik on Navy Road (Figure 4.4). Seven permafrost cores obtained from hummocks (4 cores) and interhummock troughs (3 cores) were analyzed for moisture content, excess-ice content, organic-matter content and sediment texture. The stratigraphic patterns of ice accumulation in permafrost were interpreted based upon a detailed active-layer history for the site (Mackay, 1995), which indicates a paleoactive layer with a 1981 thaw unconformity at its base. Following active-layer deepening in

Figure 4.4 The Navy Road study area (N) north of Inuvik. Active-layer and permafrost cores were obtained from hummocky terrain colonized by a black spruce and alder forest.

the 1970's and formation of the thaw unconformity, the active-layer rapidly thinned during the 1980's (4 cm a^{-1}) in association with the proliferation of alder (Mackay, 1995). The active layer remained relatively stable throughout the 1990's until the time of coring.

It was also hypothesized that vertical moisture migration beneath a degrading permafrost table is limited to the top 10 cm of permafrost. This hypothesis was tested by obtaining a core in summer 2000 then removing the top of the hummock the following spring (2001) causing the near-surface permafrost to degrade. Ground-ice samples were collected during late summer 2001 and compared with patterns of ice accumulation prior to disturbance, to determine the vertical extent of moisture migration beneath a degrading permafrost table.

Moisture migration into permafrost following aggradation of the permafrost table was also investigated by comparing vertical patterns of ice enrichment from variously aged surfaces in the Mackenzie Delta. Ice content was determined in near-surface permafrost on aggrading-point bars. These ice contents were compared with that of older permafrost from a corresponding stratigraphic level in core profiles from elevated surfaces on the Delta plain. In addition, active-layer cores obtained late in winter and early in the thaw season were analyzed to determine if moisture from the upper active layer, saturated during spring thaw, had been redistributed into the dry sediments of the underlying frozen active layer.

Objective 3. To examine temporal aspects of ground-ice development.

Initial rates of ice accumulation in near-surface permafrost may be on the order of 1 to 10 mm yr⁻¹ (Burn, 1988), but Mackay has suggested (pers. comm. with Burn, 1988) that these rates may decrease with time as near-surface permafrost becomes enriched with ice. It was hypothesized that rates of ice accumulation into near-surface permafrost decrease over time. This hypothesis was tested at the Navy Road study area (Figure 4.4) where the age of permafrost varied with respect to depth below the ground surface. Determination of excess water from permafrost intervals of different age could provide estimates of moisture accumulation at the biennial, decadal and millennial time scales. For example, organic materials added to a hummock top following coring in early summer 2000 caused the permafrost table to rise by about 10 cm. Core samples were retrieved from this interval in fall 2001 providing an estimate of moisture accumulation in permafrost over a biennial time scale. The 1981 thaw unconformity (Mackay, 1995) provided a stratigraphic marker at the decadal time scale. Estimates of water content from the interval between this unconformity and base of the contemporary active layer indicated ice enrichment over decadal time scales. Ice content and organic content of permafrost below the 1981 thaw unconformity were analyzed to investigate the presence of the Holocene thaw unconformity which could provide a stratigraphic marker at the millennial time scale.

Objective 4. To determine if the near-surface ice-rich zone of permafrost is a sink for soluble nutrients.

Although the near-surface ice-rich zone is widespread, little is known about its geochemical characteristics. Laboratory experiments indicate that water and soluble ions may move together along temperature gradients in freezing and frozen soils (Cary and Mayland, 1972; Qui *et al.*, 1988), therefore it was hypothesized that near-surface ice-rich permafrost is solute-rich with respect to the overlying active-layer.

To test this hypothesis, the geochemical characteristics of soil pore water from the active layer and underlying permafrost were evaluated at 30 sites from a range of geomorphic environments in Mackenzie Delta and adjacent upland terrain near Inuvik. Soluble cation concentrations in the top 30 to 50 cm of permafrost were estimated for individual cores and compared with the concentrations in mineral soils of the overlying active layer. In individual cores, the vertical pattern of soluble ion accumulation was compared with the pattern of ice enrichment.

Objective 5. To determine the effect of aggradational ice development and degradation on spruce forest structure.

Subarctic “drunken” forests underlain by permafrost are characterized by randomly leaning spruce trees (Zoltai, 1975a). On hummocky terrain, annual heave and settlement of the active layer can shift trees, causing them to lean (Zoltai and Tarnocai, 1974). However, in even-aged stands of fire origin, tilting is initiated when trees are between 50 and 100 years old (Zoltai, 1975b). This observation suggests

that aggradational ice development associated with post-fire active-layer thinning causes the overlying ground to heave and the trees to tilt (Zoltai, 1975b).

It was hypothesized that aggradational ice development results in tree tilting and evolution of the “drunken” forest. To test this hypothesis, the relation between near-surface ground-ice characteristics and tree lean was investigated at 18 sites throughout eastern Mackenzie Delta (Figure 4.5). The tilt characteristics of six to ten spruce trees immediately surrounding the respective permafrost core sites were measured. Patterns of ice enrichment observed in the permafrost cores were corroborated by observations of ground ice at adjacent cutbanks.

Cryogenic soil movements can cause spruce trees to tilt (Zoltai and Tarnocai, 1974), inducing compression-wood growth on the leaning side of coniferous trees (Scurfield, 1973). The effect of aggradational ice development and ground displacements on tree movements in hummocky terrain was also investigated at the Navy Road study site (Figure 4.4). Reaction-wood records from 30 black spruce growing on earth hummocks were analyzed to determine how the trees responded to periods of active-layer deepening and degradation of ground ice and periods of active-layer thinning and aggradational ice development. The patterns of reaction-wood growth in the black spruce trees were interpreted in the context of a thirty-year record of active-layer development at that site (Mackay, 1995).

Figure 4.5 Eastern Mackenzie Delta and tree tilt study sites. Stars indicate the sampling sites.

4.2 Field and laboratory methods

The following methods were used to obtain and analyze active-layer and near-surface permafrost samples and to investigate relations between ground-ice development and spruce-forest structure.

4.2.1 Core retrieval and estimation of the permafrost table

Soil cores were extracted from near-surface permafrost terrain from depths of up to 3.5 m using a 5 cm diameter CRREL core barrel. The core samples were sectioned in either 5 or 10 cm intervals, logged, double bagged, and returned to a laboratory in Inuvik. Samples were kept in a cool dark place until they could be analyzed, usually within a week of collection.

If samples were obtained prior to maximum thaw, the sites were revisited later in the season to determine maximum active-layer development. Depth of thaw was determined using a steel probe pushed down to depth of refusal. However, depth to refusal could differ by a few centimetres from that of the 0°C isotherm because of the high unfrozen water content of some soils (Mackay, 1995; Burn, 1999). During summer 1999, the accuracy of probing was calibrated by inserting a temperature probe into the ground with five calibrated thermistors, spaced at 10 cm intervals, with the first thermistor at the bottom of the probe. Active-layer probing was accurate to within a few centimetres. During coring, the location of the permafrost table was estimated based on an increase in ground resistance to drilling and an increase in visible ice content (Shur, 1988). Thaw depths obtained by probing at

core sites in late August and early September indicated that estimates of active-layer depth made during coring were reliable.

4.2.2 Analysis of core samples

4.2.2.1 *Gravimetric moisture content, excess ice content and pore-water extractions*

Laboratory analyses were conducted on the soil samples to determine soil moisture and excess ice content. Soil pore water was extracted and pore-water electrical conductivity, dissolved solids and pore-water chemistry were determined.

Permafrost and active-layer samples were thawed and homogenized, poured into beakers, weighed and allowed to settle. Volumes of sediment and supernatant water were recorded in order to estimate excess ice content of the samples using:

$$[5.1] \text{ Excess ice (\%)} = [(W*1.09)/(S+(W*1.09))]*100$$

where W is the volume of supernatant water (cm³), multiplied by 1.09 to estimate the equivalent volume of ice, and S is the volume of saturated sediment (cm³).

Supernatant water for geochemical analysis was extracted with a syringe and filtered through a 0.45 µm millipore cellulose filter. If the sample was not supersaturated, an equal weight of distilled, deionized water was added. The sample was then mixed thoroughly, re-weighed and the sediment was allowed to settle for twelve hours. The supernatant water was then extracted and filtered. The

remainder of the sample was dried for 24 hours at 105°C so that field moisture content and the moisture content at extraction could be determined.

4.2.2.2 *Geochemical analyses of soil pore water*

The electrical conductivity and total dissolved solids of soil pore water were measured immediately after filtering, with a Yellow Springs Instruments model 30 conductivity/salinity probe, calibrated daily against a NaCl reference solution. Water soluble Ca^{++} , Mg^{++} , Na^+ , and K^+ extracted from the soil samples were analyzed using an atomic adsorption spectrophotometer at Simon Fraser University. Since the moisture content at extraction was determined for each soil sample, soluble ion concentrations in soil-water extractions could be converted to milliequivalents per unit weight of dry soil ($\text{meq } 100\text{g}^{-1}$) using the following calculation:

$$[5.2] \text{SSCC}_a = \text{SCC} * [\text{MCE}/100]$$

$$\text{SSCC}_b = \text{SSCC}_a/10$$

Where SCC is the soluble cation concentration in the extracted pore water (meq l^{-1}) and MCE is the gravimetric moisture content of the soil sample at extraction (%). SSCC_a is the soil soluble cation concentration ($\text{meq } 1000\text{g}^{-1}$ dry soil) and SSCC_b is the soil soluble cation concentration ($\text{meq } 100\text{g}^{-1}$).

Presentation of soluble cation concentrations per unit weight of dry soil ($\text{meq } 100\text{g}^{-1}$) facilitated comparison of solute concentration between core samples of varying moisture contents.

4.2.2.3 Organic-matter content, grain-size and unfrozen moisture content determinations

A semi-quantitative method of soil organic-matter content of oven-dried samples was obtained by the loss-on-ignition method (Sheldrick, 1984). Selected core sections were subsampled (5 g), crushed, pre-weighed and burned at 450°C for 45 minutes. The samples were left to cool and were then re-weighed so that organic-matter content could be estimated.

Soil texture was determined on representative samples from various terrain types in the Mackenzie Delta and from samples obtained at Navy Road. Organic matter was removed from samples mechanically and remaining organic materials were burned using hydrogen peroxide. Samples were oven-dried and weighed. The gravel and sand fractions were determined by wet sieving and proportions of silt and clay were evaluated using the pipette method (Sheldrick, 1984).

Unfrozen moisture content characteristic curves for the different types of soils encountered in the study area were determined in the laboratory by time domain reflectometry (TDR). The TDR trace represented a thawing cycle of remoulded samples. The measurement instrument has resolution of 2.5% by volume in unfrozen water (Patterson and Smith, 1981).

4.2.3 Site surveys

Detailed topographic surveys of point-bar transects were referenced to the mean low water level for August. To compare vertical patterns of ice accumulation in sediments of the Mackenzie Delta, moisture content profiles from different alluvial surfaces were plotted against elevation above mean August water levels for sites in the southern, central and outer Mackenzie Delta (see Figure 3.1). Land-surface elevation was surveyed at each core location with respect to water level in the adjacent channel, and active-layer depths were subtracted from the surveyed elevation at each site. Water levels at Inuvik and Tununuk on the survey dates were obtained so that permafrost table elevations at the respective sites could be corrected to the August mean water level. The difference between mean August stage and that on the surveyed date at the Inuvik station was applied to sites in the central and southern Delta. Stage data at Tununuk Point were used to correct elevations at study sites in the outer (northern) Delta.

Gill (1972, 1973), Pearce *et al.* (1988) and Pearce (1993, 1994) described the vegetation at the Delta sites and Mackay (1995) described the vegetation at Navy Road. Site vegetation was described during the field visits.

4.2.4 Spruce forest characteristics, Mackenzie Delta and reaction-wood series, Navy Road, Inuvik

Forest characteristics including tree tilt, orientation of lean, diameter at breast height and tree height were recorded for six to ten white spruce trees at 18

permafrost core sites in the Mackenzie Delta. The angle of tree lean was measured by placing an electronic tilt meter against the leaning side of a tree trunk at belt height. Spruce forest community was identified based on descriptions by Pearce *et al.* (1988).

At Navy Road, 30 black spruce trees growing on mud hummocks were cut and disks were obtained so that the rings could be studied. Stress caused by tilting of the tree can induce growth of compression wood on the leaning side of coniferous trees (Scurfield, 1973). Growth rings provide the age of the tree, while reaction-wood rings indicate the time of perturbation, and the duration and orientation of leaning (Zoltai and Tarnocai, 1974; Zoltai, 1975a). Tree disks were sanded and rings were counted under 10 to 20x magnification.

CHAPTER 5. NEAR-SURFACE GROUND ICE IN SEDIMENTS OF THE MACKENZIE DELTA

A near-surface ice-rich zone is commonly encountered in permafrost terrain, but it is not ubiquitous (Mackay, 1972). The range of environmental conditions in an alluvial environment with permafrost provides opportunities to investigate factors that influence development of near-surface segregated ground ice. Previous work on point-bar environments (Gill, 1972), vegetation succession (Gill, 1973, 1975; Pearce *et al.*, (1988), sedimentation (Pearce 1993, 1994) and permafrost (Smith, 1975, 1976) in Mackenzie Delta provide a context for interpreting the near-surface ground ice conditions determined in this study. The first part of this chapter examines the spatial distribution of near-surface ground ice in sediments of the Mackenzie Delta. The environmental conditions associated with ice-poor and ice-rich permafrost are identified and relations between vegetation community and near-surface ground-ice characteristics are discussed. In addition, the vertical patterns of ice enrichment in sediments of the Mackenzie Delta are examined.

5.1 Results

5.1.1 Ground-ice in point-bar and spruce-forest environments of the Mackenzie Delta.

Near-surface permafrost with low gravimetric water content (Table 5.1, Figure 5.1a) was observed in sediments of aggrading point bars (see Figure 3.4). The alluvium at the willow sites was sandy silt and silty sand (Figure 5.2), and maximum thaw depths were between 100 and 140 cm. The low frost susceptibility of this alluvium was indicated by the soil's unfrozen moisture content characteristic curve which demonstrated that less than 5% of pore water remained unfrozen at temperatures below -1°C (Figure 5.3). Cores obtained at willow sites in late winter 2000 indicated that the base of the active layer was desiccated (Figure 5.4a). Saturated thawed soils were commonly observed overlying desiccated frozen active-layer sediments when coring point-bar sites in early summer (Figure 5.4b). For example, at a willow site in mid-July, the frozen active layer remained dry even though overlying thawed sediments had been saturated during snowmelt, and again in early summer by river water that was introduced to the site with a water pump (Figure 5.4b).

Alder colonizes poorly drained meander-scroll depressions on aggrading point bars (Gill, 1972). The soils were sandy silts (Figure 5.2) and active-layer thickness was between 90 and 120 cm (Figure 5.1b). The mean moisture content of near-surface permafrost was 45%, but excess-ice content was low (Table 5.1).

	Willow	Alder	Spruce/alder- bearberry	Spruce/ feathermoss	Spruce/ crowberry- lichen	Lakeside alder (southern and central), sedge (northern Delta)
Totals						
Mean moisture content (%)	34	45	51	108	139	134
<i>Core sections</i>	60	15	25	60	30	30
Excess ice (%)	0	4	5	29	42	32
<i>Cores</i>	12	3	5	11	6	6

Table 5.1 Mean gravimetric moisture content and excess ice content of upper 50 cm of permafrost, point-bar willow and alder communities, spruce forests and lakeside alder and sedge communities, southern, central and outer (northern) Mackenzie Delta. Data is in Appendix 1.

Figure 5.1 The gravimetric water content profiles for active layer and near-surface permafrost for individual cores from different vegetation communities in central Mackenzie Delta. Cores (a) C-w2 to (d) C-sf2 were obtained from a point-bar transect in central Mackenzie Delta (see Figure 4.2) and (e) C-sc5 was obtained about 20 km north of Inuvik (see Figure 4.5). (a) C-w2, point-bar willow community; (b) C-a2, point-bar alder community; (c) C-sab1, spruce/alder-bearberry forest; (d) C-sf2, spruce/feathermoss forest; and (e) C-sc5, spruce/crowberry-lichen forest. The solid line marks the permafrost table.

Figure 5.2 Texture of soils from aggrading point-bar and spruce vegetation communities in the central Mackenzie Delta. Sand (■); silt (◇); clay (△). Values are the average of samples obtained from individual cores to a depth of 1.5 m from the respective vegetation communities. The number of core samples analyzed is indicated in brackets. Core samples from the willow/horsetail (C-w1), willow (C-w2), alder (C-a2), spruce/alder (C-sab1) and spruce/feathermoss (C-sf5) communities are from the transect in central Mackenzie Delta (see Figure 4.2) (Gill's camp) and the core from the spruce/crowberry-lichen site (C-sc4) was obtained north of Inuvik (D1) (see Figure 3.1).

Figure 5.3 Unfrozen water content characteristic curves for Mackenzie Delta alluvium. Curves are three-point moving averages. Sandy silt (♦) (sand 46%, silt 48%, clay 6%) from a point-bar willow site (C-w2); silty loam (◇) (sand 6%, silt 80%, clay 14%) from a spruce/crowberry lichen site (C-sc4); and silty loam (○) (sand 4%, silt 77%, clay 19%) from a lakeshore alder community (C-la2).

Figure 5.4 Gravimetric water content of cores from two point-bar willow communities, Mackenzie Delta. (a) Frozen active layer, willow site (C-w7) near Inuvik, April 2000; and (b) partially thawed active layer, willow site (C-w2), central Mackenzie Delta study site (see Figure 4.2), early July 1999. The solid line indicates the permafrost table and the dashed line indicates the depth of thaw at the time of coring.

At spruce/alder-bearberry sites, the silty loam formed static cryosolic soils (Figure 5.2) in which an active layer 80 to 120 cm deep developed (Figure 5.1c). The zone of ice-enrichment immediately beneath the base of the active layer was between 30 and 40 cm thick (Figure 5.1c) with thin, hair-like ice lenses, but excess ice content was less than 10% (Table 5.1).

At spruce/feathermoss sites, a shallow active layer consisted of surface organic material between 10 and 30 cm thick underlain by silty loam (Figure 5.2). The underlying ice-rich zone ranged up to 2 m thick in the southern Mackenzie Delta but was less than 1 m thick in the central Delta (Figure 5.1d). Mean gravimetric moisture content in the top 50 cm of permafrost was 108% and mean excess ice content was 29% (Table 5.1). Ice content declined with depth (Figure 5.1d).

Several shallow cores were obtained from spruce/crowberry-lichen communities on Delta surfaces raised above the level of spring flooding. In the central Delta, a thin active layer (< 70 cm) was composed of clayey silts (Figure 5.2) overlain by organic materials. The moisture content characteristic curve indicated that at temperatures below -1°C, the unfrozen moisture content of the clayey silts was only 2 to 3% higher than in sandy silts (Figure 5.3). An ice-rich zone consisting of segregated ice lenses immediately beneath the base of the active layer was encountered at all spruce/crowberry-lichen sites (Table 5.1). The mean gravimetric moisture content of near-surface permafrost at spruce/crowberry-lichen sites was 139% and mean excess ice content in the top 50 cm of permafrost was 42% (Table 5.1).

5.1.2 Ground-ice characteristics in other Delta environments

In the central Delta, alder colonizes wetland and lakeshore sites while sedges occupy similar environments north of treeline (Mackay, 1963). Such low-lying surfaces are frequently inundated by floodwaters but rates of sedimentation are low because of their distance from major stream channels (Pearce, 1993, 1994). Active-layer and near-surface sediments were fine-grained silty loam with clay contents between 15 and 20%. The unfrozen moisture content of these sediments was only slightly greater than that of the silty loam from spruce/crowberry-lichen sites (Figure 5.3). A surface organic layer was present at all sites and active-layer depths ranged from 40 to 90 cm. Near-surface permafrost was ice-rich (Figure 5.5) with a mean excess ice content of 32% (Table 5.1). The ice-rich interval extended to depths of about 2 m. At depth, sediments were darker grey with abundant shells and lower organic contents than at shallower depth. Although the deeper sediments were ice-rich, the ice-content was generally less than in the near-surface (Figure 5.5).

Figure 5.5 Gravimetric water content of active layer and near-surface permafrost, lakeshore and sedge wetland site, Mackenzie Delta. a) C-la3, lakeshore alder community, Gill's camp, central Mackenzie Delta (see Figure 4.2); b) O-s2, lakeshore sedge community, outer Mackenzie Delta (see Figure 4.3). The solid line marks the permafrost table.

5.2 Distribution of near-surface ground ice in sediments of the Mackenzie

Delta

In the Mackenzie Delta, terrestrial flooding and sedimentation influence vegetation growth (Gill, 1972, 1973; Pearce *et al.*, 1988), active-layer conditions (Smith, 1975) and ground thermal regimes (Smith, 1976). Table 5.1 indicates near-surface ground-ice conditions associated with the different Delta vegetation communities. Analyses of variance indicate that the mean moisture content in near-surface permafrost was significantly different ($P < 0.01$) between all vegetation communities with the exception of the alder and spruce/alder bearberry communities and the spruce/crowberry-lichen and lakeshore alder/sedge communities.

In summary, point-bar (willow and alder) and spruce/alder-bearberry communities, characterized by sandy silts (Figure 5.2), thick active layers (< 90 cm) (Figure 5.1a, b, c), frequent flooding (Gill, 1973) and high sedimentation rates (2 to 20 cm a^{-1}) (Gill, 1973; Pearce, 1993, 1994), were underlain by ice-poor permafrost. In contrast, high ice-content permafrost was encountered beneath lakeside alder and sedge communities. These latter sites were poorly drained and sedimentation rates ($> 2 \text{ cm a}^{-1}$) were lower than at point-bar sites of similar elevation but adjacent to stream channels (Pearce, 1993, 1994). The sediments at lakeside alder sites were fine-grained silts with the highest unfrozen water contents of any terrestrial sediments obtained from the Delta (Figure 5.3).

Spruce/feathermoss communities were underlain by medium to high ice contents with a mean excess ice content in near-surface permafrost of 29% (Table 5.1).

Near-surface excess ice content at spruce/crowberry lichen sites was 42% and active layers were thin. The spruce/feathermoss and spruce/crowberry-lichen communities are characterized by low sedimentation rates (0 to 0.2 cm a⁻¹) (Pearce *et al.*, 1988).

5.3 Discussion: Ground-ice conditions with reference to soil and ground-thermal conditions, Mackenzie Delta

An ice-rich zone is widespread, but not ubiquitous in the near-surface permafrost of the Mackenzie Delta (Table 5.1). Factors influencing development of near-surface segregated ice were investigated in the Delta where a range of vegetation, soil and thermal conditions was associated with different ground-ice characteristics (Section 5.1). Localized pockets of injection ice, formed by the intrusion of water, under pressure, into confining material, may also comprise ice in near-surface permafrost (Mackay, 1972, 2000), but tabular masses of pure ice were not encountered during core sampling in this study.

The low frost-susceptibility of sandy silts in aggrading point-bar environments (Figure 5.3) may contribute to the ice-poor permafrost at these sites (Table 5.1). The low permeability of the sandy silts was indicated by frozen sediments at the base of the active layer, which remained dry despite being overlain by saturated thawed soils (Figure 5.4b). The rapid freezing of soil water that infiltrated into the

frozen sandy silts may have blocked pore spaces and restricted additional water movement deeper into the soil profile.

The thermal regime of near-surface permafrost at aggrading point bars may also inhibit development of ice-rich permafrost. Perfect and Williams (1980) demonstrated that at temperatures just below 0°C, the rates of thermally-driven water flow through frozen silts decreased by an order of magnitude when the temperature gradient was dropped from 0.1 to 0.05°C cm⁻¹. A gentle temperature gradient decreases the suction gradient that develops in freezing soils (Penner and Goodrich, 1981). The late-summer temperature gradient in near-surface permafrost at a willow site in the outer Delta was less than 0.02°C cm⁻¹ (Figure 5.6a). At this site, near-surface permafrost temperatures at 2 m depth ranged from -0.9°C in summer to -1.4°C in late winter (Figure 5.6a). This variation in temperature corresponds to a difference of only 4% in unfrozen water content for sandy silt alluvium (Figure 5.3). Therefore, the seasonal variation in soil hydrologic conditions would be minimal, decreasing the seasonal difference between the summer and winter moisture flux.

Active-layer moisture conditions at point-bar willow communities also restrict ice accumulation in underlying permafrost. Warm soil temperatures associated with willow communities can promote evaporation (Smith, 1975), but the very low summer soil moisture contents (Gill, 1972) can also be attributed to transpiration loss from dense willow cover (Blanken and Rouse, 1995). Dry soils in summer limit potential moisture movement into permafrost, whereas in winter, under-

Figure 5.6 Ground-temperature profiles for April (♦) and August (◊) at sites in the Mackenzie Delta. The data for Site a) were provided by L.D. Dyke, GSC and ground temperatures for Sites b) and c) are from Smith, 1973, Appendix 4; Site 4 and Site 5.

saturated active-layer conditions favour upward vapour flux (Williams and Smith, 1989) that may contribute to desiccation at the base of the active layer (Figure 5.4a) and moisture loss from permafrost. At these sites, deposition of alluvium and surface aggradation resulting in an upward rise in the permafrost table may incorporate dry active layer soils into the top of permafrost.

The ice content at point-bar alder communities was slightly higher than at the willow sites (Table 5.1), which can in part be attributed to wetter active-layer soils associated with alder sites. The ground-thermal regime may also promote development of near-surface ground ice. For example, the summer temperature gradient in near-surface permafrost at an alder community at the central Delta study site (CD) was slightly steeper than at a willow site in the outer Delta (Figure 5.6) (Smith, 1973). Furthermore, the seasonal variation in near-surface permafrost temperatures at the alder site indicates that the difference in seasonal moisture flux at the top of permafrost was greater than at the willow site (Figure 5.6).

Spruce/alder-bearberry sites, also underlain by ice-poor permafrost, were characterized by a dry active layer. Sedimentation rates in these communities are between 1 to 2 cm a⁻¹ (Pearce *et al.*, 1988; Pearce, 1993, 1994), causing rapid upward aggradation of the permafrost table and providing little time for ice accumulation in near-surface permafrost. Spruce/feathermoss forests replace spruce/alder-bearberry communities in association with surface aggradation and a decline in flood frequency and a decrease in sedimentation rate (Pearce *et al.*, 1988; Pearce, 1993, 1994). Although soil texture was similar between these two forest

types (Figure 5.2), high ice content at spruce/feathermoss sites indicates that in the Delta, sediment texture alone does not inhibit ice accumulation in permafrost.

At spruce/feathermoss sites, a decrease in sedimentation and the establishment of moss cover modifies the ground-thermal regime, causing the active layer to thin (Smith, 1975). The cooler ground temperatures may affect soil moisture conditions by decreasing evaporative losses, and gradual active-layer thinning provides a longer period of time for ice to accumulate at the top of permafrost (Cheng, 1983).

The thermal-regime at spruce/feathermoss sites may also contribute to moisture accumulation in near-surface permafrost. High soil suctions and a rapid decrease in permeability with depth are conditions that favour the growth of ice lenses (Penner and Goodrich, 1981), and thus, the development of near-surface ground ice. At the spruce/feathermoss site, a steep temperature gradient characterizes near-surface ground-thermal conditions in summer (Figure 5.6c) (Smith, 1973). Furthermore, near-surface permafrost temperatures vary seasonally from just below 0.0°C to about -9.0°C (Figure 5.6c) indicating that unfrozen water content and permeability of near-surface permafrost soil at these sites are several orders of magnitude greater in summer than in winter.

The highest near-surface ice contents were observed at spruce/crowberry-lichen sites. Spruce/crowberry-lichen forests grow on the oldest Delta surfaces elevated above the level of regular flooding and upward aggradation of the permafrost table due to sedimentation is very low (Pearce *et al.*, 1988). The permafrost table may slowly continue to aggrade due to the accumulation of surface organic material, but

at these sites, moisture enrichment in the top of permafrost has proceeded over several centuries. Ice accumulation over a longer time period can explain the comparatively high near-surface ice contents at these sites with respect to the spruce/feathermoss sites (Table 5.1).

Fine-grained, frost-susceptible sediments (Figure 5.3) and low rates of sedimentation characterized lakeside alder and sedge communities (Pearce, 1993, 1994) underlain by high ice-content permafrost (Table 5.1). These environments were poorly drained and remained saturated at the end of the thaw season. Ground temperatures are not known, but the low rates of sedimentation, soil and moisture conditions that characterize these sites are conducive to the development of near-surface aggradational ice.

5.4 Factors affecting ice accumulation in sediments of the Mackenzie Delta

Previous studies have indicated the importance of grain-size in determining ice accumulation in alluvial sediments (Shur and Jorgenson, 1998). In permafrost samples from the Mackenzie Delta, the moisture content decreased as sand content increased (Figure 5.7a) and increased with silt content (Figure 5.7b) which suggests that ice-rich permafrost was associated with fine-grained alluvium. However, the considerable scatter in all relations in Figure 5.7 indicates that ice-rich sediments may be associated with a range of sediment texture. For example, the sand and clay content of ice-rich sediments (> 80% moisture content) were between 5 and 45%,

Figure 5.7 Comparison of gravimetric water content with (a) sand, (b) silt, and (c) clay content of soil samples from near-surface permafrost, Mackenzie Delta.

and 5 and 24%, respectively. Furthermore, the frost susceptibility of several samples of Delta alluvium of different texture were generally similar (Figure 5.3). These data suggest that factors other than grain-size influence the development of aggradational ice in sediments of the Delta.

In addition to sediment texture, environmental conditions associated with ice-rich permafrost in the Mackenzie Delta are:

1) Low sedimentation rates.

Sedimentation rates of 0.1 to 2 cm a⁻¹ characterize elevated sites on the Delta plain colonized by spruce/feathermoss, as well as low-lying lakeshore and wetland sites (Pearce, 1993, 1994; Pearce *et al.*, 1988). Spruce/crowberry-lichen sites are rarely if ever inundated under current flooding regimes (Pearce *et al.*, 1988). The direct influence of low sedimentation is to slow the rate of permafrost aggradation and increase the time available for ice accumulation into near-surface permafrost.

2) Moist soil conditions.

Moisture migration from the base of the active layer into permafrost is influenced by active-layer soil moisture. For example, saturated soils at the base of the active layer during late summer provide moisture for migration into the top of permafrost. Lakeside alder and sedge communities underlain by ice-rich permafrost are the most mesic of terrestrial environments in the Delta, while point-bar willow communities underlain by ice-poor permafrost characterize the driest terrain in the Delta (Gill, 1972).

3) Cold permafrost temperatures.

Summer temperature gradients at sites with cold permafrost tend to be steeper than at sites with warm permafrost. A steep temperature gradient favours moisture migration in freezing soils and the development of ice lenses (Perfect and Williams, 1980). Greater seasonal variation in near-surface permafrost temperatures and greater variation in soil hydrologic conditions were associated with sites underlain by cold permafrost. This condition enhances the seasonal imbalance in moisture flux and promotes annual accumulation of moisture at the top of permafrost (Cheng, 1983).

5.5 Vertical distribution of ground ice in sediments of the Mackenzie Delta

To compare vertical patterns of ground ice in sediments of the Mackenzie Delta, moisture content profiles from different alluvial surfaces were plotted against elevation above mean August water levels for sites in the southern, central and outer Delta (Figure 5.8a, b, c). Ice-poor permafrost with moisture content between 30 and 50% characterized relatively low-lying sites on aggrading point bars throughout the Delta (Figure 5.8a, b, c). In the central Delta, the permafrost at spruce/crowberry-lichen sites was ice-rich with moisture contents in near-surface permafrost that exceeded 150% (Figure 5.8b). At spruce/crowberry-lichen and spruce/feathermoss sites, high near-surface ground-ice contents declined with depth to similar amounts measured in near-surface permafrost on aggrading point bars (Figure 5.8b).

Figure 5.8 Gravimetric water content of near-surface permafrost with respect to elevation above mean August water level for

(a) southern, (b) central and (c) outer (northern) Mackenzie Delta. Samples from an individual core are indicated by symbol.

Core profiles from: aggrading point bars ($\blacklozenge$, $\blacktriangle$); spruce/alder-bearberry ($+$, $\square$); spruce/feathermoss and spruce/crowberry-lichen ($\blacktriangledown$, $\diamond$, $\times$); and lakeside alder and sedge ($\blacksquare$, $\circ$) communities.

In the southern Delta, moisture content profiles indicated a similar pattern with low moisture contents in near-surface permafrost of aggrading point bars and high ice contents in the near-surface permafrost at elevated sites (Figure 5.8a). High ice content in near-surface permafrost at elevated sites declined with depth (Figure 5.8a, b). Although cores in excess of 2.5 m depth were not obtained, Pihlainen and Johnston (1953) collected deep cores from alluvial sediments at Aklavik. Their data indicates that the top 2 to 3 m of permafrost may be ice-rich and is underlain by ice-poor sediments similar to those in near-surface permafrost on aggrading point bars (Figure 5.8a, b, c).

While ice content in near-surface permafrost may be similar between spruce/feathermoss communities in the southern and central Delta, the thickness of the ice-rich interval may be greater in the south. The ice-rich interval in the central Delta was between 0.5 and 1 m thick in contrast to about 2 m thickness in the southern Delta (Figure 5.8a, b). Conditions that favour ice enrichment into the top of permafrost are associated with spruce/feathermoss communities (Section 5.3) and the difference in thickness of the ice-rich interval can be attributed to the greater range of levee heights at which spruce forests in the southern Delta can grow, and thus the greater range of elevations conducive to development of aggradational ice.

The decline in ice content with depth beneath spruce forests suggests that these surfaces originated from a point-bar environment (Figure 5.8b). The ice-poor sediments that were deposited in a point-bar environment remain preserved at depth in deeper cores from spruce forest sites (Figure 5.8b). Preservation of ice-poor

permafrost beneath a thin ice-rich zone at spruce/feathermoss and spruce/crowberry-lichen sites demonstrates that ice-enrichment was restricted to a zone immediately beneath the base of the active layer (Figure 5.8b).

Ground-ice profiles from lakeshore alder sites in the central Delta and wetland sedge communities in the outer Delta show that ice-rich permafrost may also develop in lower-lying terrain (Figure 5.8 b, c). Although these sites flood frequently, rates of sediment deposition are lower than on aggrading point bars at similar elevations (Pearce, 1993; 1994), the substrate is finer and the terrain is poorly drained. These environmental conditions favour development of an ice-rich zone (Section 5.3). Although deep cores from elevated surfaces did not indicate ice enrichment at depth, it is conceivable that some elevated Delta surfaces may have originated as lakeshore or wetland sites. Such terrain could be underlain by a much thicker ice-rich interval than observed beneath spruce sites, which have evolved from a point-bar environment (Figure 5.8).

CHAPTER 6. NEAR-SURFACE GROUND-ICE CHARACTERISTICS IN HUMMOCKY TERRAIN, NAVY ROAD

Near-surface ground-ice characteristics were examined in hummocky terrain approximately 2 km north of Inuvik on Navy Road (see Figure 4.4). The objective of field activities at Navy Road was to investigate the vertical extent of moisture accumulation by examining patterns of near-surface ground-ice accumulation with respect to an aggrading, degrading and static permafrost table. In addition, stratigraphic markers at the decadal and millennial time scales enabled estimation and comparison of ground-ice accumulation over various temporal scales. The moisture and excess ice content for cores presented in Chapter 6 are in Appendix 2.

6.1 Results: soils, gravimetric water content and organic-matter content

6.1.1 Soils

At Navy Road, the earth hummocks consist of clayey silt (Table 6.1) and interhummock depressions are underlain by organic-rich material. Soil textures are similar to those reported by Mackay (1980, Figure 1) for a hummock excavated at a nearby site. An unfrozen moisture content characteristic curve for a remoulded soil sample from permafrost beneath an earth hummock was determined using the TDR technique (Patterson and Smith, 1981). As temperature dropped from -0.25 to -0.75°C , the unfrozen water content decreased from about 35% to 12%, but

Depth (cm)	Sand (%)	Silt (%)	Clay (%)
20-40	9	48	43
90-100	12	48	40
120-130	8	53	39
160-170	3	47	50
200-210	30	50	20

Table 6.1 Textural distribution of sediment samples in active-layer and permafrost samples from a hummock top (Core N1), Navy Road, Inuvik. Coarser materials in some samples consisting of a few small pebbles constituted less than 5% sample weight.

remained above 5% to a temperature of -8°C (Figure 6.1). This demonstrates that in the clayey silts, considerable unfrozen moisture exists at temperatures below 0°C , and that the amount of unfrozen moisture decreases almost fourfold between -0.25 and -1.25°C . Smith (1985) obtained a freezing characteristic curve with a similar inflection point for a silty clay loam from a nearby site.

6.1.2 Moisture content

Gravimetric water content profiles from the active layer and permafrost from the middle of four hummocks were obtained by drilling (Figure 6.2). An ice-rich zone, recorded directly below the base of the active layer was 10 to 20 cm thick with moisture contents between 100 and 150%. Ice lenses in this zone were abundant. Lower moisture contents and ice-poor permafrost were observed below this upper ice-rich zone, but another increase in moisture content was present between 130 and 140 cm, and extended to a depth of 160 cm. This ice-rich interval contained ice lenses up to 1 cm thick. A second interval of lower moisture contents between 160 and 180 cm depth was underlain by a third distinct zone of ice enrichment up to 30 cm thick.

Beneath interhummock depressions, a zone of ice enrichment was observed directly below the base of the contemporary active layer (Figure 6.3). However, an increase in gravimetric water content between the active layer and near-surface permafrost was not pronounced in all cores because of the high organic-matter content of the active layer. An interval of lower moisture content 10 to 40 cm thick

Figure 6.1 Unfrozen water content characteristic curve for clayey silt from near-surface permafrost, Navy Road, Inuvik. The curve is a three-point moving average.

Figure 6.2 Profiles of gravimetric water content in the active layer and near-surface permafrost beneath hummock tops, Navy Road, Inuvik. The thick line marks the permafrost table in 2000, the thin line marks the depth of thaw in 1981 and the dashed line marks the estimated depth of a deep thaw unconformity.

Figure 6.3 Profiles of gravimetric water content in the active layer and near-surface permafrost beneath interhummock depressions, Navy Road, Inuvik. The thick line marks the permafrost table, the thin line marks the depth of thaw in 1981 and the dashed line marks the estimated depth of a deep thaw unconformity.

was observed below this upper ice-rich zone. A second ice-rich zone at the 90 cm depth in Cores N5 and N7 and at the 110 cm depth in Core N6 was 10 to 20 cm thick and contained ice lenses up to 1 cm thick (Figure 6.3). Lower moisture contents below this zone of ice enrichment extended to depths of about 150 cm in Cores N6 and N7. A third increase in moisture content was observed in all cores below the 150 cm depth (Figure 6.3).

Comparison of an undisturbed moisture content profile beneath a hummock top with the pattern of ice enrichment following active-layer deepening and thawing of near-surface permafrost indicated the vertical extent of ice-enrichment beneath a degrading permafrost table. Core N2 obtained in August 2000 described the ice-content profile of near-surface permafrost prior to disturbance (Figure 6.2 and 6.4). The remaining two cores were obtained in September 2001 following the removal of the hummock top early that spring (Figure 6.4). A thin zone of ice enrichment immediately below the thaw front was detected in both cores obtained following the degradation of the upper 20 cm of permafrost (Figure 6.4). Two to three ice lenses about 0.2 to 0.4 cm thick were observed between 115-118 cm depths in these two cores, but the moisture contents of ice-poor permafrost between 120 and 130 cm depth remained low following the degradation of overlying permafrost. The ice-rich zone below 130 cm was present before and after disturbance (Figure 6.4).

Figure 6.4 Profiles of gravimetric moisture content in near-surface permafrost beneath a hummock top, Navy Road, Inuvik (▲). Moisture content before degradation of near-surface permafrost (N2), August 2000, and (□, +) after degradation of top 20 cm of permafrost (N8 and N9), September 2001. The thick line marks the permafrost table in August 2000, the thin line marks the extent of maximum active-layer development (September 2001) following perturbation and the dashed line marks the depth of thaw in 1981. Note that the y-axis does not start at the ground surface.

Figure 6.2 – N4 and Appendix 2) were obtained from a hummock top in 2000 and 2001, respectively. After the first core was obtained, 50 cm of moss and peat were added to the hummock top causing the permafrost table to aggrade by approximately 10 cm. The first core (N4) indicated that the moisture content at the base of the frozen active layer between the depths of 90 and 100 cm was 58%. Following aggradation of the permafrost table in 2000, these depths remained perennially frozen. In late summer 2001, the moisture content of this zone was 60% (Core N4b) and although a few thin ice lenses were observed, moisture enrichment was negligible.

6.1.3 Organic-matter content

Organic-matter content was determined for core samples from a hummock (Core N1) and an interhummock trough (Core N5). Beneath the hummock, organic-matter content in the active layer ranged from 5 to 7% (Figure 6.5a). From the permafrost table to a depth of 170 cm, the organic-matter content ranged from 6% to 9%. An organic-rich horizon was encountered between 170 and 190 cm depths.

The active layer and top 10 to 30 cm of permafrost beneath interhummock depressions consisted of organic-rich materials (Figure 6.5b). Mineral soils with low organic-matter content below were underlain by a distinct organic horizon at the 170 to 180 depths (Figure 6.5b).

Figure 6.5 Soil organic-matter contents in active layer and permafrost, Navy Road, Inuvik: (a) N1 – hummock centre; (b) N5 – interhummock depression. The thick line marks the permafrost table, the thin line marks the depth of thaw in 1981 and the dashed line marks the estimated depth of a deep thaw unconformity.

6.2 Discussion

6.2.1 Thaw unconformity

Three zones of ice-enrichment were observed in permafrost profiles beneath hummocks and interhummock troughs at Navy Road (Figure 6.2, 6.3 and 6.6). A near-surface zone of ice-enrichment consisting of segregated ice lenses, immediately below the base of the present active layer was approximately 10 to 20 cm thick (Figure 6.6). Excavation of near-surface permafrost indicated a small body of injection ice beneath a hummock periphery (Figure 6.6). This ice probably resulted from the freezing of water injected into the hummock periphery by cryostatic pressure in the interhummock trough during an autumn freeze-back period.

The second increase in ice content at 130 cm depth beneath hummock tops and at 90 to 110 cm depth beneath interhummock depressions coincided with the top of permafrost during maximum active-layer development in 1981 (Mackay, 1995) (Figure 6.6). This zone of ice enrichment was 30 to 40 cm below the base of the active layer beneath hummocks and about 50 to 60 cm below the base of the active layer beneath interhummock troughs. A third ice-rich zone between 180 and 220 cm depths and 160 and 190 cm depths was observed beneath hummocks and interhummock troughs respectively (Figure 6.6). An organic horizon coincided with the top of the third zone of ice enrichment beneath hummock tops, but occurred within the third zone of ice-enrichment beneath interhummock troughs (Figure 6.3 and 6.5).

Figure 6.6 Near-surface ground ice beneath hummocky topography at Navy Road, Inuvik. Core samples beneath hummock centres and interhummock depressions indicated three zones of segregated ice accumulation. Excavation of near-surface permafrost indicated abundant segregated ice lenses and a small body of injection ice.

Ice-rich intervals immediately below the base of the contemporary permafrost table and beneath the depth of maximum thaw in 1981 are consistent with Cheng's (1983) hypothesis of segregated ice accumulation in near-surface permafrost. Preservation of the ice-rich interval below the 1981 thaw unconformity has provided a stratigraphic marker. In near-surface permafrost on Herschel Island, Yukon, a similar moisture (ice) content pattern in conjunction with a geochemical discontinuity was interpreted to represent the position of the permafrost table during maximum thaw in the early Holocene (Kokelj *et al.*, 2002).

At Navy Road, the third and deepest zone of ice enrichment suggests a thaw unconformity at the base of a paleoactive layer extending beneath hummocks to depths of about 180 cm (Figure 6.2a, b, c) and beneath interhummock depressions to depths of about 150 to 170 cm (Figure 6.3a, b, c). If the present day hummocky micro-relief is considered, shallower paleoactive-layer depths beneath interhummock depressions indicate that during maximum thaw the permafrost table was planar (Figure 6.6). Beneath hummock tops, this deep ice-rich interval was overlain by a discrete organic horizon between 170 and 190 cm depth (Figure 6.5). Organic-rich materials within the deep zone of ice enrichment beneath interhummock troughs suggest that maximum active-layer development may have slightly exceeded depth estimates based upon moisture content profiles.

There is a considerable amount of near-surface ground ice above the deep thaw unconformity. Subtracting the excess ice content from the interval between the base of the active layer and the unconformity provides an estimate of paleoactive-layer

thickness. Accounting for excess ice in near-surface permafrost, estimated thaw depths of 130 cm beneath interhummock troughs and 170 cm beneath hummock tops are greater than the maximum measured post-fire active layers of 102 cm depth for interhummock depressions and 136 cm depth for hummocks at an adjacent area which burned in 1968 (Mackay, 1995). The deep unconformity may therefore indicate deeper active-layer development during warmer climate conditions in the early Holocene (Burn, 1997) or more recently during the 'Medieval Warm Period' (Bégin *et al.*, 2000).

6.2.2 Vertical extent of moisture migration into permafrost, Navy Road

The extent of moisture movement into permafrost at Navy Road was interpreted based upon the detailed active-layer history for the site presented by Mackay (1995). Active-layer thinning beneath hummock tops from 132 cm in 1981 to 90 cm in 1992 indicated upward aggradation of the permafrost table by approximately 4 cm a⁻¹, followed by a period of stability throughout the 1990's to the time of data collection in 1999-2001. Low rates of moisture accumulation into permafrost during the period of rapid active-layer thinning in the 1980's were indicated by an ice-poor interval between the 110 cm and 130 cm depths (Figure 6.2). An ice-rich zone between 100 and 110 cm developed during a period of permafrost stability between 1991 and 2001 (Figure 6.2). The preservation of underlying ice-poor permafrost between 110 and 130 cm depths demonstrates that under equilibrium

permafrost conditions, moisture enrichment is limited to a thin zone immediately below the base of the active layer.

Moisture migration below the thawing front into ice-poor permafrost was investigated by removing a hummock top in the spring of 2001. Moisture contents in the upper 5 cm of permafrost beneath the degrading thaw-front were greater than that at the same depth prior to the disturbance (Figure 6.4), but underlying permafrost remained ice-poor despite saturated conditions at the base of the deepening active layer.

At an annual time scale, moisture accumulation ahead of a degrading permafrost table was detected only 5 cm below the thaw front, despite saturated conditions at the base of the thawing active layer. Permafrost degradation, development of a deep thaw unconformity and the maintenance of a thick active layer for at least several centuries (Burn, 1997) resulted in ice enrichment 10 to 30 cm beneath the depth of maximum thaw. An ice-rich interval of similar thickness was observed beneath the Holocene paleoactive layer on Herschel Island (Kokelj *et al.*, 2002, Figure 5). Accounting for excess ice in these ice-rich intervals, moisture enrichment into originally ice-poor permafrost was limited to the top 5 to 20 cm. Preservation of moisture-content patterns associated with the deep thaw unconformity (Figure 6.2) indicates that over centennial to millennial time scales, moisture accumulation at these depths has been negligible despite the fine-grained, frost-susceptible nature of the sediments (Table 6.1, Figure 6.1).

These results support the interpretation that atmospherically-derived tritium in the top 10 to 30 cm of ice-rich permafrost is a result of temperature-induced water migration since the mid-1950's (Michel and Fritz, 1978; Chizhov *et al.*, 1983; Burn and Michel, 1988). However, the inferred deeper movement of water into permafrost over annual time scale using TDR (Cheng, 1983), and tritium profiles where the surface of permafrost terrain was irrigated with tritium (Burn and Michel, 1988) is difficult to reconcile. It is possible that the tritiated water moved into permafrost down thermal contraction cracks that develop each winter (Burn and Michel, 1988), or that elevated levels were the result of contamination from shallower depths in the core sample.

Moisture movement deeper into near-surface permafrost is restricted because in summer, temperature, and therefore the frozen soil's hydraulic conductivity, declines with depth. Downward migration of moisture is also inhibited by the presence of ice lenses that decrease the permeability of frozen soil at temperatures just below 0°C (Burt and Williams, 1976). Under equilibrium permafrost conditions, net accumulation of moisture at depth (> 30 cm) should be negligible since seasonal differences in the unfrozen moisture content and hydraulic conductivity of near-surface permafrost soils decrease with depth where the amplitude of seasonal temperature variation is small.

6.2.3 Moisture migration over various temporal scales

Stratigraphic markers that identified decadal and millennial thaw unconformities and the manipulation of surface vegetation resulting in the aggradation of the permafrost table enabled estimates and comparison of moisture accumulation in permafrost over various time scales.

Seasonal rates of moisture accumulation were estimated based on samples taken from the base of the frozen active layer and recently aggraded permafrost. The moisture content at the base of the frozen active layer in mineral soils was between 45 and 60%. The moisture content in two-year-old permafrost was 60%. Some thin ice lenses were observed, but samples did not yield excess water upon thawing.

Moisture accumulation over decadal time periods was estimated based on permafrost samples obtained between the base of the present active layer and the 1981 thaw unconformity. The mean amount of excess water in this interval was 29 mm (Table 6.2). Following aggradation of the permafrost table, a thin ice-rich zone developed during a decade of equilibrium permafrost conditions in the 1990's. The mean annual rate of moisture enrichment over the last two decades could be estimated by dividing time (20 years) by the excess water content in this interval of near-surface ice-rich permafrost. Therefore, mean annual rate of water accumulation into recently aggraded, previously ice-poor permafrost was approximately 1.5 mm a^{-1} .

Moisture accumulation in permafrost at the millennial time scale was estimated based on samples obtained between the contemporary permafrost table and the deep

Time scale and number of cores	Estimated minimum and maximum amount of excess water (mm)	Mean amount of excess water (mm)	Estimated rates of moisture accumulation in permafrost (mm a^{-1})
Decadal (1981-2001) N = 7	12 to 45	29	1.5
Millennial (1000 to 10,000 BP – present) N = 5	108 to 222	151	0.015 to 0.15

Table 6.2 Estimated amounts of excess water and rates of moisture accumulation in near-surface permafrost over decadal and millennial time scales, Navy Road, Inuvik. Due to uncertainty in the age of the deep thaw unconformity, moisture accumulation rates were estimated over 1000 years (Medieval Warm Period) and 10,000 years (Early Holocene Climate Optimum).

thaw unconformity. Under contemporary climatic conditions, it is unlikely that fire-induced disturbance caused active-layer thaw depths to exceed those in 1981, so the deep unconformity probably formed during a period of warmer climate during the 'Medieval Warm Period' (Bégin *et al.*, 2000) or the early Holocene Climate Optimum (Burn, 1997). The mean amount of excess water above the deep unconformity in five cores was 151 mm (Table 6.2). If the deep unconformity dated to the early Holocene the mean rate of moisture accumulation into aggrading permafrost since that time has been about 0.015 mm a^{-1} , but if active-layers development during the 'Medieval Warm Period' approached the depth of the unconformity, a mean rate of moisture accumulation of 0.15 mm a^{-1} is estimated (Table 6.2).

The mean annual rates of moisture accumulation into 20-year-old permafrost, on the order of one to two mm a^{-1} (Table 6.2), complement Burn's (1988) estimates of 10 mm a^{-1} . Burn's measurements using soil analogue tubes were conducted only over a single season and Mackay suggested (pers. comm., Burn, 1988) that these rates might decrease over time. Thus, initial rates of moisture accumulation into young permafrost may not represent long-term rates. For example, at Navy Road the estimated rate of moisture accumulation at the millennial scale was two to three orders of magnitude lower than the annual rate measured by Burn (1988) and one to two orders of magnitude lower than accumulation rates into two-decade-old permafrost under a static permafrost table. If the deep thaw unconformity at Navy Road is of early Holocene age, moisture accumulation above this unconformity was

about one order of magnitude lower than Burn's (1986) estimate of moisture accumulation for that same time interval at a site in central Yukon.

Vegetation succession, disturbance or climate change influence active-layer development and affect the amount of aggradational ice measured above unconformities that developed under warmer climate conditions. For example, ice-poor permafrost developed when upward aggradation of the permafrost table was rapid, but during the following decade of static permafrost table conditions, the top 10 to 20 cm of permafrost became enriched with ice (Figure 6.2 and 6.3). Ice-rich intervals 10 to 30 cm in thickness beneath paleoactive layers demonstrate that near-surface ground ice may develop in the absence of an aggrading permafrost table, but indicate the limited vertical extent of ice accumulation. Therefore, seasonal variation in active-layer thaw and very gradual aggradation of the permafrost table favour the development of a thick ice-rich zone, but disturbance such as fire causing active-layer deepening and degradation of the ice-rich zone can release water that has accumulated in near-surface permafrost over decades to several millennia.

CHAPTER 7. NUTRIENT ACCUMULATION IN NEAR-SURFACE

PERMAFROST

Although the near-surface ice-rich zone is widespread, little is known about its geochemical characteristics. Laboratory experiments indicate that water and soluble ions move together along thermally-induced suction gradients during the freezing of fine-grained soils (Cary and Mayland, 1972; Qui *et al.*, 1988), therefore ice accumulation in near-surface permafrost may be accompanied by solute enrichment. This chapter describes the chemical characteristics of active-layer and near-surface sediments in the Mackenzie Delta and the Inuvik area. The process of solute accumulation in near-surface permafrost is examined and water and nutrient sequestration in the ice-rich zone are considered in the context of fire ecology.

7.1 Results

7.1.1 Geochemical characteristics of active layer and near-surface permafrost in the Mackenzie Delta region

Soluble Ca^{++} and Mg^{++} were the dominant cations found in active-layer and near-surface permafrost soils of the central Mackenzie Delta (Table 7.1). In core samples, mean total soluble cation concentrations in mineral soils of the active layer showed a decreasing trend with distance from the stream channel. Total soluble cation concentrations in the active layer ranged from $2.06 \text{ meq } 100\text{g}^{-1}$ dry soil on an aggrading surface colonized by willows, to $0.28 \text{ meq } 100\text{g}^{-1}$ at a spruce/crowberry-lichen forest above the level of regular flooding (Table 7.1). Total soluble cation

Site	Near-surface moisture content (%)	Soluble cations (meq 100g ⁻¹ dry soil)									
		AL	P	AL	P	AL	P	AL	P	AL	P
		Ca ⁺⁺	Ca ⁺⁺	Mg ⁺⁺	Mg ⁺⁺	Na ⁺	Na ⁺	K ⁺	K ⁺	Totals	totals
C-w2 - Point-bar willow community	34	1.05	2.28	0.93	1.49	0.06	0.05	0.01	0.01	2.06	3.83
C-a2 – Point-bar alder community	57	0.75	1.28	0.72	0.78	0.07	0.08	0.01	0.01	1.54	2.15
C-sab1 - Spruce/alder-bearberry forest	56	0.26	1.22	0.26	0.78	0.10	0.08	0.01	0.01	0.64	2.09
C-sf4 – Spruce/Feathermoss forest	72	0.24	0.50	0.16	0.67	0.09	0.06	0.01	0.01	0.49	1.24
C-sc5 – Spruce/Crowberry-lichen forest	156	0.15	1.18	0.15	0.98	0.11	0.06	0.01	0.02	0.28	2.25

Table 7.1 Gravimetric moisture content of near-surface permafrost and mean soluble cation concentrations at the base of the active layer (AL) and top 50 cm of permafrost (P) for several sites along a point-bar transect, in central Mackenzie Delta (see Figure 4.2) and from a spruce/crowberry-lichen community north of Inuvik (see Figure 4.5) raised above the level of regular flooding.

concentrations in permafrost were between 1.4 and 8 times greater than in sediments of the overlying active layer and the relative solute enrichment of permafrost was greatest at the ice-rich, spruce/crowberry-lichen site (Table 7.1).

Soluble Ca^{++} and Mg^{++} were the dominant cations in active layer and near-surface permafrost on low-lying terrain in the outer Delta (Table 7.2). However, unlike permafrost in the southern and central Delta, the relative abundance of soluble Na^+ increased with depth in the core profile. For example, at a site in low-lying terrain between 1 and 1.5 m a.s.l., Na^+ concentrations increased from 0.04 meq 100g^{-1} in near-surface permafrost to 1.18 meq 100g^{-1} at the 200-210 cm depth (Table 7.2). Sodium was the dominant cation below 110 cm depth.

On elevated terrain adjacent to the Mackenzie Delta at the Navy Road site, total soluble cation concentrations in the active layer of hummock tops were generally below 0.2 meq 100g^{-1} (Table 7.3). Soluble Ca^{++} was a dominant cation but Na^+ and K^+ were also present in relatively high amounts (Table 7.3). Only traces of soluble Mg^{++} were present in the active layer and near-surface permafrost, but concentrations increased by one order of magnitude at a depth of 140 cm and remained high deeper in the profile.

Depth (cm)	Soluble cations (meq 100g ⁻¹ dry soil)				Total cations
	Ca ⁺⁺	Mg ⁺⁺	Na ⁺	K ⁺	
30-40	0.41	0.18	0.07	0.01	0.66
50-60	0.34	0.25	0.04	0.00	0.63
80-90	0.61	0.35	0.06	0.01	1.03
110-120	0.35	0.48	0.46	0.01	1.30
140-150	0.05	0.13	0.41	0.00	0.59
170-180	0.04	0.16	0.67	0.01	0.87
200-210	0.13	0.23	1.18	0.01	1.55
230-240	0.51	0.24	1.17	0.01	1.93

Table 7.2 Active-layer and permafrost soluble cation concentrations in low-lying terrain adjacent to a large lake (Core O-s2), outer Mackenzie Delta (see Figure 4.3).

The permafrost table is at 40 cm.

Depth (cm)	Soluble cations (meq 100g ⁻¹ dry soil)				Totals
	Ca ⁺⁺	Mg ⁺⁺	Na ⁺	K ⁺	
20-40	0.06	0.02	0.04	0.05	0.16
40-60	0.05	0.01	0.04	0.04	0.13
60-70	0.04	0.01	0.02	0.04	0.10
70-80	0.09	0.01	0.04	0.05	0.19
80-90	0.08	0.01	0.02	0.04	0.15
90-100	0.05	0.02	0.04	0.06	0.16
100-110	0.17	0.02	0.05	0.06	0.30
110-120	0.11	0.01	0.02	0.05	0.19
120-130	0.09	0.02	0.05	0.07	0.23
130-140	0.09	0.05	0.05	0.07	0.27
140-150	0.32	0.20	0.32	0.16	0.99
150-160	0.19	0.13	0.10	0.06	0.48
160-170	0.19	0.11	0.06	0.03	0.40
170-180	0.21	0.18	0.06	0.02	0.47
180-190	0.33	0.24	0.10	0.04	0.72
190-200	0.39	0.13	0.07	0.16	0.75
200-210	0.30	0.15	0.05	0.16	0.66

Table 7.3 Soluble cation concentrations, active layer and permafrost, hummock top (Core N2), Navy Road, Inuvik. The single line at 100 cm depth indicates the permafrost table, the single line at 130 cm depth indicates depth of maximum thaw in 1981, and the thick line indicates a deep thaw unconformity (see Section 6.2.1).

7.1.2 Soluble cation enrichment of near-surface permafrost

Mean total soluble cation concentrations in mineral soils of the active layer were plotted against concentrations in the top 30 to 50 cm of underlying permafrost for sites throughout the Mackenzie Delta and the Inuvik region (Figure 7.1). Active-layer total soluble cation concentrations ranged from less than $0.03 \text{ meq } 100\text{g}^{-1}$ dry soil at a mature black spruce forest site in hummocky upland terrain to more than $2.0 \text{ meq } 100\text{g}^{-1}$ at a spruce/alder-bearberry site in Delta alluvium. Minimum and maximum soluble cation concentrations in permafrost of 0.09 and $4.9 \text{ meq } 100\text{g}^{-1}$ were also measured at these two sites (Figure 7.1). Although concentrations varied from site to site, permafrost was solute-rich with respect to the active layer at 29 of 30 sites (Figure 7.1).

Enrichment ratios were calculated for all sites presented in Figure 7.1 by dividing the mean total soluble cation concentration in the top 30 to 50 cm of permafrost by the mean concentration in mineral soils at the base of the active layer (Appendix 3a). Permafrost enrichment ratios for specific sites ranged from 0.7 to 8, but on average, the cation concentrations in permafrost were 2.8 times greater than in the overlying active layer. The data were divided into two groups based on the ice content of near-surface permafrost. At sites underlain by medium to high ice-content permafrost (excess ice content $>10\%$), mean cation enrichment was 3.3, while the mean enrichment ratio for ice-poor sites was only 1.6. An analysis of variance indicated that the mean nutrient enrichment of permafrost between low and high ice content sites was significantly different ($F_{1,28} = 9.49$, $P < 0.01$).

Figure 7.1 Mean soluble cation concentrations in mineral soils of the active layer and mean concentration in the top 30 to 50 cm of underlying permafrost for cores obtained in the Mackenzie Delta region. Cores were obtained during summers 1999 to 2001, from sites in alluvial sediments of Mackenzie Delta ($\diamond$) and upland sites in the Inuvik region (X). For data see Appendix 3a.

7.1.3 Soluble nutrient accumulation in the near-surface ice-rich zone

In near-surface permafrost, solute accumulation generally corresponded with zones of ice-enrichment. For example, at a spruce/alder-bearberry site in the Delta, the soluble cation concentration at the base of the active layer was $0.57 \text{ meq } 100\text{g}^{-1}$ but increased in near-surface permafrost to a maximum of $2.68 \text{ meq } 100\text{g}^{-1}$ at 110 to 120 cm depth (Figure 7.2a). The zone of maximum ice-enrichment was also recorded at the 110 to 120 cm interval (Figure 7.2b). Soluble cation concentrations and moisture content both decreased below 120 cm depth (Figure 7.2).

Similar patterns were observed at sites with ice-rich permafrost. At a spruce/crowberry-lichen site in the Delta, an ice-rich zone was observed immediately beneath the base of the active layer between depths of 60 cm and 110 cm (Figure 7.3b). The soluble cation concentration in mineral soils at the base of the active layer was $0.24 \text{ meq } 100\text{g}^{-1}$ and increased by an order of magnitude to $3.5 \text{ meq } 100\text{g}^{-1}$ in near-surface permafrost (Figure 7.3a). The zones of maximum soluble cation concentration and ice-enrichment coincided at the depth of 100 to 110 cm (Figure 7.3).

At Navy Road, zones of moisture enrichment in near-surface permafrost have developed beneath the contemporary permafrost table and the paleo-thaw unconformities (see Chapter 6). Cation accumulation (Table 7.3 and Figure 7.4a) corresponded with the three respective ice-rich intervals (Figure 7.4b). For example, soluble cation concentrations increased from $0.15 \text{ meq } 100\text{g}^{-1}$ at the base of the active layer to $0.3 \text{ meq } 100\text{g}^{-1}$ in an ice-rich interval at 100 to 110 cm depth.

Figure 7.2 (a) Total soluble cation concentrations and (b) gravimetric moisture content for active layer and near-surface permafrost, spruce/alder-bearberry community (Core C-sab1), Mackenzie Delta (see Figure 4.2). The solid line indicates the permafrost table. Note that the y-axis does not start at the ground surface.

Figure 7.3 (a) Total soluble cation concentrations and (b) gravimetric moisture content for mineral soils of the active layer and near-surface permafrost, spruce/crowberry-lichen community (Core C-sc5), Mackenzie Delta (see Figure 4.5). The solid line indicates the permafrost table. Note that the y-axis does not start at the ground surface.

Figure 7.4 (a) Total soluble cation concentrations and (b) gravimetric moisture content for active layer and near-surface permafrost, hummock top (Core N2) black spruce forest, Navy Road, Inuvik. The thick line marks the permafrost table, the thin line marks the depth of thaw in 1981 and the dashed line marks the estimated depth of maximum thaw. Note that the y-axis does not start at the ground surface.

Lower soluble cation concentrations extended to a depth of 140 cm where concentrations increased from 0.26 to 1.0 meq 100g⁻¹. The increase in soluble cations corresponded with an ice-rich interval between the 130 and 160 cm depths. A third increase in solute concentrations corresponded with the deepest zone of ice-enrichment between the 180 and 210 cm depth.

Several permafrost cores from a spruce/crowberry-lichen site in the Mackenzie Delta were sectioned in 5 cm rather than 10 cm intervals. The moisture content and total soluble cation profiles of two cores are shown in Figures 7.5 and 7.6. Total cation concentrations at the base of the active layer of the respective cores were 0.25 (Figure 7.5a) and 0.56 meq 100g⁻¹ (Figure 7.6a). Soluble cation concentrations increased in near-surface permafrost and maximum concentrations were 2.87 (Figure 7.5a) and 3.12 meq 100g⁻¹ (Figure 7.6a), respectively. In both cores, the zone of maximum solute concentration was 10 to 15 cm below the base of the active layer and immediately above the zone of maximum ice-enrichment (Figures 7.5 and 7.6). The moisture-rich intervals had excess ice contents between 50 and 70%.

7.2 Discussion

7.2.1 The geochemistry of near-surface permafrost in the Mackenzie Delta region

The dominance of soluble Ca⁺⁺ and Mg⁺⁺ in alluvial sediments reflects the carbonate bedrock and tills of the Mackenzie Basin from where the floodwaters and sediments were derived (Carson *et al.*, 1998). During spring flood of Mackenzie Delta, Ca⁺⁺ concentrations in a stream channel were 28.1 mg l⁻¹ (Lesack *et al.*,

Figure 7.5 (a) Total soluble cations and (b) gravimetric water content for active layer and permafrost, spruce/crowberry-lichen community (Core C-sc7), Mackenzie Delta (Site D1, see Figure 3.1). The solid line indicates the permafrost table. Note that the y-axis does not start at the ground surface and that the core was sectioned in 5 cm intervals.

Figure 7.6 (a) Total soluble cations and (b) gravimetric water content for active layer and permafrost, spruce/crowberry-lichen community (Core C-sc8), Mackenzie Delta (Site D1, see Figure 3.1). The solid line indicates the permafrost table. Note that the y-axis does not start at the ground surface and that core was sectioned in 5 cm intervals.

1991). Concentrations of Mg^{++} , Na^+ and K^+ were 6.2, 4.5 and 1.02 mg l^{-1} respectively. On Delta surfaces, soluble cation concentrations were influenced by the frequency of inundation by solute-rich floodwaters. For example, active-layer concentrations in alluvial soils at a willow site on an aggrading point bar were an order of magnitude greater than the concentrations at a spruce/crowberry-lichen site above the level of regular flooding (Table 7.1).

Table 7.1 also indicates that solute concentrations in near-surface permafrost beneath elevated surfaces of the Delta plain are similar to concentrations in the active layer of lower-lying surfaces that experience frequent flooding and surface aggradation. Permafrost aggradation may preserve high soluble cation concentrations at depth and leaching of soluble ions from active-layer sediments at elevated sites can contribute to the contrast between active-layer and permafrost soluble cation concentrations (Kokelj *et al.*, 2002).

In the outer Delta, soluble Ca^{++} and Mg^{++} were the dominant cations in the active layer and near-surface permafrost, but at several sites, the relative abundance of Na^+ increased with depth (Table 7.2). The grey colour and the dominance of soluble Na^+ indicated that these sediments were deposited in a shallow-water marine environment. Sodium chloride can also be introduced to low-lying terrain in the outer Delta by incursion of dilute seawater during summer storm surges (Jenner and Hill, 1991). As low-lying terrain in the outer Delta aggrades, flooding and sediment deposition are dominated by high-water events associated with spring melt. The pore-water chemistry of near-surface sediments thus reflects fresh-water deposition,

while the Na^+ in underlying sediments remains immobilized by permafrost (Table 7.2).

Nutrient-rich alluvial sediments generally possessed higher soil soluble cation concentrations than sediments of upland terrain in the Inuvik region where surficial deposits consist of carbonate tills and colluvium (Figure 7.1). At Navy Road, active-layer soils were nutrient poor and Ca^{++} was a dominant cation while Mg^{++} was present in lowest relative abundance (Table 7.3).

At Delta sites, the relative proportions of the different cations present in the active layer were generally similar to those in underlying permafrost (Table 7.1). Exceptions were sites where the nature of deposition had changed over time. For example, in the outer Delta, geochemical profiles indicated a shift from brackish or marine to fresh-water deposition (Table 7.2). The relative abundance of ions may also differ above and below stratigraphic discontinuities such as thaw unconformities (Kokelj *et al.*, 2002). This pattern was suggested at Navy Road, where the absolute amount and relative abundance of soluble Mg^{++} increased below the depth of maximum active-layer development in 1981 (Table 7.3).

7.2.2 The ice-rich zone as sink for soluble nutrients

Although the abundance of soluble cations in the soil varied throughout the study region, concentrations in permafrost were greater than in the overlying active layer at virtually all sites (Figure 7.1). Removal of salts from seasonally-thawed sediments by surface and subsurface water movement can lead to a contrast in the

geochemical characteristics between the active layer and underlying permafrost (Leibman and Streletskaya, 1997; Kokelj and Lewkowicz, 1999; Kokelj *et al.*, 2002). Furthermore, in freezing soils, water and soluble ions move together along thermal gradients (Cary and Mayland, 1972; Qui *et al.*, 1988; Chuvilin *et al.*, 1998), thus near-surface permafrost may become enriched with ice as well as with soluble materials (Figures 7.2 to 7.4).

Ions transported to a freezing front by convection are excluded from the ice phase and accumulate on the warm side of the developing ice lenses (Hallet, 1978). The degree of exclusion from the ice phase is a function of the freezing rate and the solute concentration at the ice-water interface (Hallet, 1978). With respect to aggradational ice development, solutes may accumulate on the warm side of (above) thick segregated ice lenses. However, Baker and Osterkamp (1988) found that the efficacy of solute exclusion was reduced during upward freezing because high interfacial solute concentrations may develop in the absence of gravity-driven brine drainage.

Soluble cation and moisture content profiles from near-surface permafrost were consistent with laboratory and theoretical descriptions of water and ionic accumulation during freezing of fine-grained soils (Hallet, 1978; Qui *et al.*, 1988; Chuvilin *et al.*, 1998). Near-surface permafrost was nutrient-rich with respect to the base of the active layer at sites throughout the study area and solute enrichment was greater at sites with ice-rich permafrost. Correspondence between soluble cation and moisture content profiles suggests that convective transport contributed to the

nutrient enrichment of near-surface permafrost (Figure 7.2 to 7.4). High resolution sampling of ice-rich permafrost demonstrated that cations were concentrated immediately above ice-rich intervals, suggesting that soluble ions drawn downward to the freezing front by convection were excluded from the growing ice lenses (Figure 7.5).

The data demonstrate that moisture and solute movement into near-surface permafrost are linked, and that these processes, in conjunction with lateral leaching of solutes from the active layer (Leibman and Streletskaya, 1997; Kokelj and Lewkowicz, 1999; Kokelj *et al.*, 2002), contribute to the contrast between active-layer and permafrost geochemical characteristics. Correspondence between moisture and soluble cation profiles indicate that near-surface aggradational ice is a sink for soluble nutrients. The ice-rich zone is commonly encountered in permafrost terrain (Mackay, 1972), therefore nutrient enrichment in near-surface permafrost may be widespread. Where the permafrost table has aggraded in response to sedimentation, vegetation succession or climate change, the ice-rich zone could exceed one metre in thickness and may constitute a significant pool of moisture and nutrients that could be made available to plants upon thawing.

7.2.3 Post-fire release of soluble nutrients from permafrost

Forest fires can enrich soils by releasing nutrients sequestered in biomass (Wien and Bliss, 1973; Viereck, 1983). Warmer soil temperatures following the destruction of surface organic cover can increase microbial activity and

decomposition rates (MacLean *et al.*, 1983; Van Cleve and Viereck, 1983). It has also been recognized that moisture and nutrients in near-surface permafrost can be released into the bottom of the active layer during periods of active-layer deepening and thaw of ice-rich permafrost (Mackay, 1995).

Post-fire soluble cation and moisture release from near-surface permafrost were estimated based on cryogeochemical data obtained from two spruce forest sites in the Mackenzie Delta region. Site D1 is a spruce/crowberry-lichen community in Delta alluvium, approximately 15 km north of Inuvik (see Figure 3.1). The second site is a black spruce woodland in hummocky terrain adjacent to the Delta at Navy Road (N) (Figure 3.1). Error in these calculations is derived from spatial variation in soluble cation amounts in the soil, and the magnitude of post-fire permafrost degradation, which would vary depending on the intensity of the fire. The purpose of this calculation however, was to provide order of magnitude estimates for comparison with the amount of nutrients in litter and standing crops that could be released upon total combustion of treeline forest communities in the Inuvik region (Wien and Bliss, 1973).

Thawed soil dry bulk densities of 1.3 and 1.4 g cm⁻³ were based on three samples collected from the base of the active layer at both the Delta (D1) and hummocky upland (N) sites. Post-fire active-layer deepening was estimated using data collected in the Inuvik area after an intense fire in 1968 (Mackay, 1995; Figure 2- Site 3). Following fire, the increase in active-layer depth was estimated to be 25 cm in year one, 12 cm in year two, 8 cm in each of years three and four, and 4 cm in

year five. Mean values for soluble cation concentrations in near-surface permafrost were derived from cores obtained from the two respective sites. Mean concentrations of specific cations in near-surface permafrost were converted from milliequivalents to grams for comparison with data reported by Wien and Bliss (1973) expressed as g m^{-2} .

Since the dry bulk density of soil at the base of the active layer was determined, the weight of soil added to the active layer per m^2 (SW) given a known amount of active-layer deepening can be calculated using:

$$[7.1] \text{ SW (g) = DBD} \times V$$

where DBD is the dry bulk density of thawed soil at the base of the active layer (g m^{-3}) and V is soil volume (m^3) for a given amount of active-layer deepening.

The amount of soluble cations contributed to the base of the active layer under 1 m^2 of the ground surface can then be estimated using:

$$[7.2] \text{ Soluble cations (g) per metre square = SW} \times \text{CC}$$

where CC is the mean concentration of a specific cation in near-surface permafrost (g g^{-1} dry soil).

In the first summer following fire, an estimated 1.87 to 43.16 g m^{-2} of soluble Ca^{++} could be released into the bottom of the active layer at Navy Road and the Delta site respectively (Table 7.4). The estimated release of K^+ , Mg^{++} and Na^+ associated with 25 cm of active-layer deepening would exceed the estimated nutrient pool in litter and standing crops of treeline forest communities by as much as one order of magnitude (Table 7.4). Five years after the fire, active-layer

Site	Year after fire and amount of active- layer deepening (cm)	Cations released from degrading permafrost (g m ⁻²)				Moisture released from degrading permafrost (cm)
		Ca ⁺⁺	Mg ⁺⁺	Na ⁺	K ⁺	
N – Navy Road	1, 25 cm	1.87	0.31	3.72	5.87	5.3
	2, 12 cm	0.90	0.15	1.79	2.82	2.5
	3, 8 cm	0.60	0.10	1.19	1.88	1.7
	4, 8 cm	0.60	0.10	1.19	1.88	1.7
	5, 4 cm	0.30	0.05	0.60	0.94	0.8
5 year total	57 cm	4.27	0.70	8.49	13.38	12.0
D1 – Mackenzie Delta	1, 25 cm	43.16	31.53	4.59	4.03	14.4
	2, 12 cm	20.71	15.13	2.20	1.93	6.9
	3, 8 cm	13.81	10.09	1.47	1.29	4.6
	4, 8 cm	13.81	10.09	1.47	1.29	4.6
	5, 4 cm	6.90	5.04	0.73	0.64	2.3
5 year total	57 cm	98.40	71.89	10.47	9.19	32.8
Litter and Standing crop on unburned areas, Nutrients (g m ⁻²)						
		Ca ⁺⁺	Mg ⁺⁺	Na ⁺	K ⁺	
Caribou Hills**		2.72	0.68	0.15	1.87	
Inuvik**		1.87	0.42	0.01	1.46	

** Data from Wein and Bliss, 1973.

Table 7.4 Estimated amounts of soluble cations and moisture contributed to the base of the active layer following degradation of near-surface permafrost. For location of sites see Figure 3.1. Data from Wein and Bliss (1973) on nutrient release following burning of subarctic forests in the Inuvik area are presented for comparison. Estimates of nutrient and moisture release from permafrost are based on mean soluble cation concentrations and excess ice content in the top 30 to 50 cm of permafrost from three cores obtained at the two respective sites (see Appendix 3b).

deepening could continue to release soluble Ca^{++} , Mg^{++} , K^+ and Na^+ to the bottom of the active layer in amounts comparable with those that would have been initially released through the combustion of biomass (Table 7.4). It should be noted that the nutrient quantity sequestered in spruce forests of the Mackenzie Delta is probably much greater than in the biomass of treeline communities in the Inuvik region.

Table 7.4 also indicates that the nature of soluble cation additions to the active layer from degrading permafrost will vary locally, depending on the geochemistry of the near-surface permafrost. This suggests that vegetation responses to permafrost degradation may be at least partly dependant on the geochemistry of the underlying permafrost.

The thaw of ice-rich permafrost by deepening of the active layer also contributes water to the bottom of the active layer. In one summer, this amount may equal or exceed the entire summer's rainfall (Mackay, 1995). For example, excess ice in the top 50 cm of permafrost was estimated to be 39% at the spruce/crowberry-lichen forest (D1) in Mackenzie Delta, and 19% in hummocky terrain at Navy Road (Appendix 3b). At the Delta site, deepening of the active layer by 25 cm would thaw approximately 41 cm of permafrost, melting about 16 cm of ice releasing about 14 cm of water to the bottom of the active layer. A similar amount of active-layer deepening would release approximately 5 cm of water into the base of the active layer at the Navy Road site (Table 7.4). In contrast, mean summer rainfall for Inuvik (June to August) is approximately 10 cm (Atmospheric and Environment Service, 2002).

The release of water and soluble ions from near-surface permafrost can enhance soil conditions and promote the growth of colonizing vegetation. The thaw of ice-rich permafrost contributes moisture to the bottom of the active layer which can offset dry soil conditions typical of recently burned terrain in non-permafrost or ice-poor areas (Rouse, 1976). The amount of soluble cations that can be released into the bottom of the active layer during periods of fire or climate-induced active-layer deepening are significantly greater than the amount of nutrients sequestered in the biomass of treeline forest communities (Table 7.4). The relative magnitude of the nutrient pool in near-surface permafrost suggests that it is ecologically significant (Table 7.4).

CHAPTER 8. THE EFFECT OF AGGRADATIONAL ICE DEVELOPMENT AND DEGRADATION ON THE STRUCTURE OF SPRUCE FORESTS

This chapter investigates the effects of ground ice development on the structure of overlying spruce forests. Previous work has related tree tilting and development of the “drunken forest” to cryoturbation within the active layer (Zoltai and Tarnocai, 1974; Zoltai, 1975a), but observations by Zoltai (1975b) suggest that aggradational ice development associated with post-fire active-layer thinning may cause the overlying ground to heave and the trees to tilt. This chapter describes the relations between spruce forest structure and near-surface ground ice in alluvial sediments in the eastern Mackenzie Delta. The effect of aggradation and degradation of near-surface ground ice on black spruce growing in hummocky terrain is also examined.

8.1 “Drunken forests” and near-surface ground ice in the Mackenzie Delta

8.1.1 Ground-ice and forest characteristics

Several shallow permafrost cores were obtained from spruce forest sites in the Mackenzie Delta. Moisture and excess ice contents of the near-surface permafrost were determined. Forest characteristics, including amount and orientation of tree lean, diameter at breast height and tree height, were recorded for six to ten white spruce trees at each of the 18 core locations.

Figure 8.1 displays data gathered at 18 sites in the Mackenzie Delta, and indicates that spruce forests with tilted trees are underlain by ice-rich permafrost. At sites

Figure 8.1. Excess ice content in the top 50 cm of permafrost and angle of tree lean for 6 to 10 spruce trees at permafrost sampling sites throughout eastern Mackenzie Delta (see Figure 4.5).

where the ice content of near-surface permafrost was greater than 40%, all trees were tilted, some by more than 25°. Most trees were leaning at sites underlain by medium ice content (10 to 35% excess ice content), but maximum tilt was generally not as great as at high ice content sites. Spruce forests with straight trees were underlain by ice-poor permafrost with excess ice contents below 15%.

Figure 8.2 indicates the mean tree tilt of six to ten spruce trees surrounding each permafrost core site and the near-surface excess ice content of the ground at each site. Forest-community type was used to group these data (Pearce *et al.*, 1988). Spruce/alder-bearberry forests consisted of straight trees underlain by ice-poor permafrost (see Sections 5.1 and 5.2, Figures 8.2 and 8.3). The mean tree lean in spruce/feathermoss forests ranged from 3 to 12° and the ice content of near-surface permafrost was medium to high (Figures 8.2 and 8.4). The mean tilt of trees in spruce/crowberry-lichen woodlands was greater than 10° and near-surface permafrost was ice-rich (Figures 8.2 and 8.5).

The mean excess ice content in near-surface permafrost for spruce/alder-bearberry sites was 5%. Ice content in the top 50 cm of permafrost at spruce/feathermoss forests ranged from 15 to 43% and the mean of nine sites was 29%. The mean excess ice content in permafrost beneath spruce/crowberry-lichen woodlands was 43%. Analysis of variance indicated that the mean ice content was significantly different ($P < 0.01$) between the three respective forest communities.

Figure 8.2 Relation between excess ice content in the top 50 cm of permafrost and mean angle of tree lean at 18 sites in eastern Mackenzie Delta (see Figure 4.5).

Forest type was used to group the data: Spruce/alder-bearberry (♦), spruce/feathermoss (▲) and spruce/crowberry-lichen (●). Data are in Appendix 4.

Figure 8.3 Spruce/alder-bearberry community (Core C-sab4) adjacent to East Channel north of Inuvik, Mackenzie Delta (see Figure 4.5). The spruce trees are straight and the near-surface ice contents are low.

Figure 8.4 Spruce/feathermoss community with leaning trees (Core S-sf1), southern Mackenzie Delta (see Figure 4.5). The excess ice content of near-surface permafrost is medium to high. Arrows indicate leaning trees. Note abundant deadfall.

Figure 8.5 Spruce/crowberry-lichen forest (Core C-sc1) with randomly leaning, stunted spruce trees underlain by ice-rich permafrost, central Mackenzie Delta (see Figure 4.5).

8.1.2 Discussion

The development of near-surface ice-rich permafrost (Mackay, 1972, 1983; Cheng, 1983) may contribute to the tilting of spruce trees in the Mackenzie Delta. On frequently flooded surfaces, such as point-bar willow and spruce/alder-bearberry communities, rapid aggradation of the permafrost table and the low frost-susceptibility of sandy silts inhibit ice enrichment in permafrost (see Section 5.2 to 5.4). As these surfaces aggrade, flood frequency decreases and finer alluvium is deposited (see Section 5.1, Figure 5.2). Ice accumulation then occurs in near-surface permafrost because frost-susceptible sediments, gradual aggradation of permafrost and colder permafrost temperatures favour ice enrichment.

In the Delta, long-lived spruce trees survive aggradation of the permafrost table and development of near-surface ground ice. Trunk and root systems of these trees may become anchored in permafrost, but their ability to grow adventitious roots enables them to survive (Gill, 1975). The evolution from a planar frost table beneath aggrading surfaces, to an undulating frost table beneath older, elevated Delta surfaces (Gill, 1975) causes differential aggradation of permafrost and variability in ground-ice development. This can enhance surface micro-relief and may contribute to the gradual tilting of trees. Where ice-rich sediments underlie spruce forests, degradation of permafrost resulting from disturbance or extreme thaw events may cause differential subsidence and tree movement. In contrast, degradation of ice-poor permafrost would not effect tree lean as thaw-settlement would be minimal.

The effect of aggradational ice development or subsequent thaw-subsidence on tree lean is suggested by data in Figure 8.1 which show that forests with straight trees are underlain by ice-poor permafrost and forests with tilted trees are underlain by ice-rich permafrost. Excess ice content plotted against mean site tree tilt illustrates that forest type can be used to predict near-surface ground-ice characteristics (Figure 8.2). The relation in Figure 8.2 is useful in determining thaw-sensitive terrain in the Mackenzie Delta as forest communities are easily distinguished (Figures 8.3 to 8.5) and can be mapped using various remote-sensing techniques.

8.2 Tilting of black spruce trees (*Picea mariana*) and development and degradation of aggradational ice

Reaction-wood records from tilted black spruce trees growing on earth hummocks in the Navy Road study area were analyzed to determine if the trees had responded to recent active-layer thinning and aggradational ice development, and to the preceding period of active-layer deepening and ground-ice degradation (Mackay, 1995). The analysis of growth rings from trunk cross-sections of 30 leaning black spruce trees are presented and discussed in the context of historical active-layer data (Mackay, 1995) and near-surface ground-ice conditions (see Section 6.2.2).

8.2.1 Tree cross-section analysis

At Navy Road, the ages of the 30 black spruce trees ranged from 52 to 93 years, with a mean age of 70 years. The stand probably developed following a fire that burned through the Inuvik region more than 120 years ago (Black and Bliss, 1978; Landhausser and Wein, 1993). The absence of black spruce in the adjacent 1968 burned area indicates that several decades may pass before spruce trees re-establish.

Some of the black spruce trees had as many as five distinct sequences of reaction-wood growth punctuated by stability indicated by normal growth rings. On average, however, trees experienced two to three periods of movement. The average duration of a reaction-wood sequence was about 10 years, however two of the trees had continuously grown reaction wood since the 1950's.

The percentage of trees that produced reaction wood in a given year indicates that there were periods of greater tree movement as well as quiescence (Figure 8.6). Prior to 1950, reaction-wood rings were observed in less than 20% of the sampled trees. Percentages increased in the 1950's and peaked in 1966 when 80% of trees produced reaction wood (Figures 8.6a, 8.7a). The prevalence of reaction wood in annual growth rings declined throughout the late 1960's and early 1970's (Figure 8.6a and 8.7a). By 1976, reaction wood was observed in only 22% of annual growth rings and remained below 30% until the early 1980's. From 1982 to 1988, the percentage of trees with reaction-wood rings increased from 27% to 97% indicating extensive local tree movement (Figures 8.6a, and 8.7). The percentage of

Figure 8.6 (a) The percentage of black spruce trees that produced reaction wood in a given year, 1930 – 2000, Navy Road, Inuvik. (b) Depth of thaw for hummocks (▲) and interhummock depressions (Δ), 1968-1993, 1999-2000, black spruce forest, Navy Road, Inuvik. Active-layer data for 1968 to 1993 are from Mackay (1995, Figure 2).

A)

B)

Figure 8.7 Trunk cross-sections from leaning black spruce trees in hummocky terrain, Navy Road, Inuvik. Periods of reaction wood are indicated by black lines. In tree (A) reaction wood indicates tree movement between 1963 and 1973 and from 1988 to the time of sampling in spring 2001; and in tree (B) reaction wood indicates that tree movement was initiated in 1987.

trees that produced reaction wood remained above 80% per annum up until sampling in 2001.

8.2.2 Interpretation

At Navy Road, ground movement during the 1950's and 1960's is indicated by an increase in the number of black spruce trees that produced reaction wood (Figure 8.6a). These earth movements may be attributed to active-layer thinning, aggradational-ice development and hummock growth associated with post-fire vegetation succession. By 1968, a bowl-shaped permafrost table was established (Figure 8.6b) and hummock form was well-developed (Mackay, 1995).

Active-layer deepening, destruction of the bowl-shaped permafrost table and degradation of ice-rich permafrost during the late 1960's and early 1970's caused hummocks to spread outward and hummocky relief to decrease (Mackay, 1995). These changes in near-surface permafrost conditions and hummock morphology coincided with the cessation of reaction-wood sequences in many black spruce trees (Figures 8.6a and 8.7a, b). A decrease in hummocky relief could right some of the leaning trees growing on hummock sides and may account for the abrupt decline in reaction wood growth in some of the trees between 1968 and 1976 (Figure 8.7a).

In the 1980's, the proliferation of alder at the Navy Road site increased ground shading causing the active layer to thin (Mackay, 1995). Accentuation of the bowl-shaped permafrost table and development of near-surface aggradational ice heaved

hummock soils and increased hummocky relief (Mackay, 1995), causing trees to tilt (Figure 8.6 and 8.7).

The reaction-wood chronologies from Navy Road demonstrate that tree tilting increased during periods of permafrost aggradation and hummock growth, and that gradual permafrost degradation and outward spreading of hummocks resulted in a decrease in tree movement or straightening of the trees. These data indicate the fundamental role of aggradational ice in the evolution of earth hummocks and development of drunken forests ubiquitous to hummocky terrain in the subarctic taiga. At Navy Road, a complex interaction between climate and vegetation succession has influenced active-layer development, near-surface ground-ice conditions (see Section 6.2.2), hummock form (Mackay, 1995) and resultant tree movements.

CHAPTER 9. CONCLUSIONS

9.1 Summary

Objective 1. To examine the spatial distribution of near-surface ground ice in sediments of the Mackenzie Delta.

This research indicated that the principal factors controlling ground ice development in sediments of Mackenzie Delta are: a) Soil texture (frost-susceptibility), b) soil moisture, c) rates of permafrost table aggradation and, d) ground-thermal regime. Although these factors are discussed in the context of the Mackenzie Delta, they influence the development of aggradational ice in all permafrost environments indicating the broad applicability of these results.

The mosaic of vegetation communities in the Mackenzie Delta reflects environmental conditions of different terrestrial surfaces (Gill, 1973). Since physical factors control ground-ice development, vegetation type was closely related to ground-ice characteristics in near-surface permafrost. The results indicate that vegetation community can be used to predict the distribution of ground ice in sediments of the Mackenzie Delta. For example, point-bar environments characterized by annual sedimentation (Gill, 1973; Smith, 1975; Pearce, 1993, 1994), rapid aggradation of the permafrost table, sandy-silt alluvium, dry active-layer soil conditions (Gill, 1972) and warm permafrost temperatures (Smith, 1973) were underlain by ice-poor permafrost. Willow, alder and some poplar colonize these environments. Alluvial environments characterized by low rates of sedimentation (Pearce, 1993, 1994; Pearce *et al.*, 1988), gradual aggradation of the

permafrost table, frost-susceptible silty loam and “cold permafrost” (Smith, 1973) were underlain by medium ice-content permafrost. Spruce/feathermoss communities colonized these sites.

Conditions at spruce/crowberry-lichen sites were similar to those at other medium to high ice-content sites, but spruce woodlands occupy the oldest surfaces of the Delta plain and the higher ice-contents at these sites may be attributed to the greater age of near-surface permafrost and stability of the permafrost table. Lakeside alder and sedge communities represent the most mesic terrestrial environments in the Delta. At these ice-rich sites, there is abundant moisture for thermally-induced migration from the base of the active layer into the permafrost during late summer.

Objective 2. To investigate vertical patterns of near-surface ground-ice accumulation with respect to aggrading, degrading and static permafrost table conditions.

The vertical patterns of ice accumulation with respect to aggradation, degrading and static permafrost table conditions were investigated in hummocky terrain at Navy Road, Inuvik. Active-layer depths beneath hummocks and interhummock depressions increased during the 1970's to a maximum of 132 and 108 cm, respectively, in 1981, but thinned during the 1980's (Mackay, 1995). Mean maximum thaw depths in 1999-2002 were similar to those measured by Mackay (1995) in the early 1990's. Addition of moss to the surface of a hummock causing

the permafrost table to aggrade by 10 cm resulted in the development of low ice-content permafrost. At undisturbed sites, core profiles indicated that rapid aggradation of the permafrost table during the 1980's (2 to 4 cm a⁻¹) resulted in the development of ice-poor permafrost. During the subsequent decade of static permafrost conditions, the top 10 to 20 cm of permafrost was enriched with ice. The top of a hummock was removed so that moisture migration into permafrost below a degrading permafrost table could be investigated. Despite moist soil conditions and frost-susceptible substrate, moisture enrichment of permafrost was detected a maximum of 5 cm below the descending thaw front.

In undisturbed terrain beneath the near-surface zone of ice-enrichment, a second increase in ice-content 20 to 30 cm thick coincided with the base of the 1981 thaw unconformity. A third, deeper ice-rich interval of similar thickness, associated with an organic horizon, was interpreted to represent maximum active-layer development during the Holocene. These ice-rich intervals demonstrate that near-surface ground ice can develop in the absence of an aggrading permafrost table, but indicate the limited vertical extent of moisture migration and ice accumulation beneath the permafrost table.

The preservation of ice-poor intervals above decadal and millennial thaw unconformities confirms that long-term moisture enrichment at depths of more than 30 cm below the permafrost table has been negligible. The limited vertical extent of thermally-induced moisture migration into permafrost was also indicated by the

preservation of ice-poor point-bar deposits which underlie a thin zone of near-surface ice accumulation at spruce forest sites in the Mackenzie Delta.

The data indicate that Cheng's (1983) hypothesis of annual ice accumulation in near-surface permafrost can explain moisture enrichment of the top 10 to 20 cm of permafrost. In hummocky terrain at Navy Road, deep zones of ice-enrichment were associated with moisture accumulation in near-surface permafrost during periods of deeper active-layer development following disturbance such as forest fire or during warmer climate conditions.

Mackay (1972) suggests that an aggrading permafrost table would trap ice lenses that form at the base of the active layer during freezeback, however, upward aggradation of the permafrost table on the order of 1 to 10 cm a⁻¹ did not favour aggradational ice development. Aggradation of the permafrost table punctuated by periods of stability, enabling ice accumulation into the very top of permafrost, or very gradual upward aggradation of the permafrost table (< 2 cm a⁻¹) are required for development of a thick ice-rich zone.

In hummocky terrain adjacent to the Mackenzie Delta, variability in active-layer development over time due to vegetation succession and climate change has resulted in the widespread development of a thick (1 m) near-surface ice-rich zone. Disturbances such as fire, which cause active-layer deepening, can release water that has accumulated in near-surface permafrost over decades to several millennia.

Objective 3. To examine temporal aspects of ground-ice development.

Cheng's (1983) hypothesis suggests that annual ice accumulation in near-surface permafrost results in development of the ice-rich zone but does not indicate whether this process can be sustained as the ice-content of permafrost increases. Permafrost cores obtained from hummocky terrain at the Navy Road study area where the age of permafrost varied with respect to depth below the ground surface indicated that initial rates of ice accumulation may be high, but decrease with time as near-surface permafrost becomes enriched with ice. The addition of organic material to the ground surface causing the permafrost table beneath a hummock to aggrade by 10 cm resulted in the development of ice-poor permafrost. However, an ice-rich zone developed in two decade-old permafrost under static permafrost table conditions. The mean annual moisture enrichment was about 1.5 mm a^{-1} . Total ground-ice accumulation above the deepest (millennial) thaw unconformity indicated that long-term rates of ice accumulation were at least an order of magnitude lower than the decadal rates.

Objective 4. To determine if the near-surface ice-rich zone of permafrost is a sink for soluble nutrients.

Active-layer and permafrost pore-water geochemistry indicated that alluvial soils on frequently flooded surfaces of the Mackenzie Delta were nutrient-rich with respect to the active-layer soils at sites elevated above the level of annual flooding.

Soluble Ca^{++} and Mg^{++} were the dominant cations in near-surface sediments throughout most of the Delta. In contrast, active-layer soils at undisturbed sites in upland terrain adjacent to the Delta were generally nutrient poor.

Throughout the Mackenzie Delta region, permafrost was solute-rich relative to mineral soils at the base of the overlying active layer. The leaching of solutes from the active layer contributes to the contrast between active-layer and permafrost geochemical characteristics (Leibman and Streletskaya, 1997; Kokelj and Lewkowicz, 1999; Kokelj *et al.*, 2002). In addition, thermal gradients in summer draw water and solutes from the active layer into near-surface permafrost. At ice-rich sites, nutrient enrichment of permafrost with respect to overlying active-layer sediments was greater than at ice-poor sites. Coupled water and solute movement and accumulation in near-surface permafrost were indicated in individual core profiles by the correspondence of solute-rich intervals with zones of ice enrichment.

The ice-rich zone is commonly encountered in permafrost terrain, therefore nutrient enrichment of near-surface permafrost may be widespread. Forest fire, climate warming or anthropogenic disturbance which lead to an increase in active-layer thaw could release moisture and nutrients sequestered in near-surface permafrost which may enhance active-layer soil conditions and promote vegetation growth. However, the geochemical effect of permafrost degradation would vary spatially because of local differences in the geochemical characteristics of near-surface permafrost.

Objective 5. To determine the effect of aggradational ice development and degradation on spruce forest structure.

The effect of ground-ice development on spruce forest structure was investigated at several sites in the Mackenzie Delta. A clear, positive association between the ice content of near-surface permafrost and the extent of tree lean at sites in the Mackenzie Delta suggested that aggradation and/or degradation of ice-rich permafrost can cause spruce trees to tilt. The effect of ground-ice development on the lean of spruce trees was corroborated at a black spruce site in hummocky terrain. Reaction-wood rings in black spruce trees indicated that a decade of active-layer thinning and aggradational ice development caused earth hummocks to grow (Mackay, 1995) and more than 90 percent of sampled trees to tilt. These observations indicate that aggradational ice development or degradation can result in the development of “drunken forest”, and that forest structure may be used to predict near-surface ground ice characteristics in permafrost terrain.

9.2 Principal findings

The principal findings of this study are as follows:

1. Rapid aggradation of the permafrost table, sandy-silt alluvium, dry active-layer soil conditions and “warm” permafrost contribute to ice-poor near-surface permafrost beneath point bars of the Mackenzie Delta. In hummocky terrain (Navy Road) underlain by fine-grained, frost-susceptible sediments, aggradation

of the permafrost table at 2 to 4 cm a^{-1} resulted in development of ice-poor permafrost.

2. In the Mackenzie Delta, ice-rich sites were characterized by moist soil conditions, static or gradually aggrading permafrost table conditions and relatively cold permafrost. At an upland site in hummocky terrain, following rapid aggradation of the permafrost table, a decade of static permafrost conditions resulted in ice accumulation in the top 10 to 20 cm of permafrost.
3. Deep zones of ice accumulation up to 30 cm thick were observed beneath decadal and millennial thaw unconformities. These zones of ice enrichment demonstrate that ground ice can develop in the absence of an aggrading permafrost table and indicate that the vertical extent of ice enrichment is limited to the top 30 cm of near-surface permafrost. The vertical extent of thermally-induced moisture migration into permafrost was also indicated by the preservation of ice-poor point-bar deposits which underlie a thin near-surface ice-rich zone at spruce forest sites in the Mackenzie Delta.
4. In hummocky terrain characterized by frost-susceptible sediments, variability in active-layer development over time due to vegetation succession and climate change has resulted in the development of a thick (1 m) near-surface ice-rich zone.
5. At Navy Road, mean annual moisture enrichment into two-decade-old permafrost was about 1.5 mm a^{-1} . Total ground ice above the millennial thaw

unconformity indicates that long-term rates of ice accumulation into near-surface permafrost are at least an order of magnitude lower than the initial rates.

6. In unconsolidated materials, the principal factors controlling the development of near-surface aggradational ice are: a) soil texture (frost-susceptibility), b) soil moisture, c) rates of permafrost table aggradation, and d) ground-thermal regime.
7. Ca^{++} and Mg^{++} were dominant soluble cations in active-layer and permafrost sediments of Mackenzie Delta. Ca^{++} was also a dominant soluble cation in sediments of upland terrain near Inuvik. Active-layer soils of frequently flooded surfaces in the Mackenzie Delta were nutrient-rich with respect to terrain elevated above the level of regular flooding.
8. Permafrost was nutrient-rich in relation to the overlying active layer throughout the study region and the relative nutrient enrichment of permafrost at high ice-content sites was greater than at sites in ice-poor terrain.
9. In permafrost profiles, zones of nutrient accumulation were associated with ice-rich intervals indicating that the ice-rich zone is a sink for both water and soluble nutrients.
10. Development of near-surface aggradational ice can cause trees in overlying spruce forests to tilt. Reaction-wood rings in black spruce trees indicated a significant increase in tree movements during periods of active-layer thinning and aggradational ice development. In the Mackenzie Delta, forests with tilted trees were underlain by ice-rich permafrost and forests with straight trees were

underlain by ice-poor permafrost. Spruce forest type may be used to predict the near-surface ground-ice characteristics in sediments of the Mackenzie Delta.

CHAPTER 10. REFERENCES

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APPENDIX 1. ACTIVE-LAYER AND PERMAFROST MOISTURE AND ICE CONTENT, MACKENZIE DELTA

1a. Southern Delta (see Figure 4.1)

Key	MC = Gravimetric moisture content (%)	EI = Excess ice content (%)
	Bold = permafrost	

Core code indicates (C-sab1): a) General location, C – central Delta; b) forest type, **sab** – spruce/alder-bearberry forest; and c) core number, 1.

Forest types: **w** - willow; **a** - alder, **la** - lakeside alder, **p** - poplar, **sab** - spruce/alder-bearberry, **sf** - spruce/feathermoss, **sc** - spruce/crowberry-lichen, **s** - sedge

S-w1, Point-bar willow community

67°52'20"N, 134°09'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	31	
	20-40	33	
	40-50	30	
	50-60	18	
	60-70	17	
	70-80	14	
	80-90	16	
	90-100	19	
	100-110	21	
	110-120	22	0
	120-130	37	0
	130-140	41	0
	140-150	42	0
	150-160	33	0

S-a1, Point-bar alder community

67°52'20"N, 134°09'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	33	
	20-40	29	
	40-60	28	
	60-80	26	
	80-90	24	
	90-100	15	
	100-110	25	
	110-120	29	
	120-130	36	0
	130-140	37	0
	140-150	37	0
	150-160	34	0
	160-170	41	0
	170-180	37	0
	180-190	36	0

S-p1, Point-bar poplar community

67°52'20"N, 134°09'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	53	
	20-40	44	
	40-60	35	
	60-70	43	
	70-80	38	
	80-90	50	0
	90-100	66	6
	100-110	57	7
	110-120	56	0
	120-130	48	2
	130-140	47	0
	140-150	43	0
	150-160	40	0
	160-170	42	0

S-sab1, Spruce/alder-bearberry community

67°52'20"N, 134°09'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	35	
	20-40	47	
	40-50	28	
	50-60	14	
	60-70	14	
	70-80	18	
	80-90	38	0
	90-100	41	0
	100-110	50	5
	110-120	52	3
	120-130	52	5
	130-140	45	7
	140-150	48	5
	150-160	58	5
	160-170	56	8
	170-180	41	0
	180-190	43	0
	190-200	45	0

S-sf1, Spruce/feathermoss community
67°52'20"N, 134°09'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	58	
	20-40	69	
	40-50	35	
	50-60	60	
	60-70	125	21
	70-80	120	25
	80-90	129	30
	90-100	160	41
	100-110	125	37
	110-120	224	52
	120-130	205	45
	130-140	148	33
	140-150	152	28
	150-160	137	34
	160-170	103	30
	170-180	213	24
	180-190	172	19
	190-200	181	18
	200-210	110	18
	210-220	85	15

S-sf2, Spruce/feathermoss community
67°52'20"N, 134°09'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	69	
	20-40	82	
	40-50	61	
	50-60	103	0
	60-70	137	0
	70-80	164	44
	80-90	160	42
	90-100	154	36
	100-110	134	26

S-sf3, Spruce/feathermoss community
67°48'00"N, 134°04'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	34	
	20-40	40	
	40-60	26	
	60-75	41	
	75-80	36	
	80-90	34	
	90-100	77	4
	100-110	78	8
	110-120	96	16
	120-130	126	30
	130-140	144	16
	140-150	117	24
	150-160	89	27
	160-170	127	29
	170-180	103	25
	180-190	143	36
	190-200	91	18

S-sf4, Spruce/feathermoss community
67°48'00"N, 134°04'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	39	
	20-40	36	
	40-60	35	
	60-80	41	
	80-90	48	
	90-100	76	21
	100-110	76	7
	110-120	122	32
	120-130	108	35
	130-140	126	37
	140-150	80	21

1b. Central Delta

(Gill's Camp) (see Figure 4.2)

Key	MC = Gravimetric moisture content (%)
	EI = Excess ice content (%)
	Bold = permafrost

C-w1, Point-bar willow/horsetail community

68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	0-10	27	
	10-20	33	
	20-30	33	
	30-40	36	
	40-50	28	
	50-60	26	
	60-70	31	
	70-80	32	
	80-90	30	
	90-100	32	
	100-110	39	0
	110-120	29	0
	120-130	29	0
	130-140	28	0
	140-150	27	0
	150-160	36	0
	160-170	30	0
	170-180	30	0
	180-190	30	0
	190-200	30	0
	200-210	30	0

C-w2, Point-bar willow community

68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	0-23	32	
	23-39	28	
	39-60	23	
	60-80	23	
	80-100	10	
	100-120	20	
	120-130	14	
	130-140	22	0
	140-150	21	0
	150-160	37	0
	160-170	36	0
	170-180	39	0
	180-190	35	0
	190-200	32	0

C-w3, Point-bar willow community

68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	0-23	36	
	20-40	31	
	40-60	33	
	60-80	30	
	80-100	28	
	100-110	18	
	110-120	20	
	120-130	36	0
	130-140	32	0
	140-150	33	0
	150-160	33	0
	160-170	28	0
	170-180	32	0
	180-190	42	0
	190-200	32	0

C-w4, Point-bar willow community

68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	0-20	43	
	22-39	27	
	39-60	24	
	60-80	29	
	80-100	31	
	100-110	32	0
	110-120	31	0
	120-130	37	0
	130-140	43	0
	140-150	35	0
	150-160	36	0
	160-170	37	0
	170-180	34	0

C-w5, Point-bar willow community
68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	0-25	32	
	25-40	34	
	40-60	32	
	60-80	33	
	80-100	31	
	100-110	29	
	110-120	23	
	120-130	24	
	130-140	22	0
	140-150	23	0
	150-160	32	0
	160-170	37	0
	170-180	34	0
	180-190	35	0
	190-200	35	0

C-w6, Point-bar willow community
68°43'00"N, 134°16'30"W

Aug-99	Depth (cm)	MC	EI
	0-25	36	
	25-40	45	
	40-60	37	
	60-80	32	
	80-90	35	
	90-100	30	
	100-110	32	0
	110-120	37	0
	120-130	47	9
	130-140	47	6
	140-150	36	2
	150-160	38	0
	160-170	41	0
	170-180	35	0
	180-190	40	0
	190-200	41	0

C-a1, Point-bar alder community
68°43'00"N, 134°16'30"W

Aug-99	Depth (cm)	MC	EI
	20-40	53	
	40-60	32	
	60-80	35	
	80-90	22	
	90-100	20	
	100-110	37	
	110-120	41	0
	120-130	48	5
	130-140	41	0
	140-150	40	0
	150-160	35	0
	160-170	32	0
	170-180	32	0
	180-190	35	0
	190-200	35	0

C-a2, Point-bar alder community
68°43'00"N, 134°16'30"W

Aug-99	Depth (cm)	MC	EI
	0-26	48	
	26-40	40	
	40-60	40	
	60-70	36	
	70-80	19	
	80-86	22	
	86-90	41	
	90-100	48	0
	100-110	56	6
	110-120	70	13
	120-130	57	18
	130-140	51	13
	140-150	52	12
	150-160	42	11
	160-170	40	0
	170-180	39	0
	180-190	40	0
	190-200	39	0

C-la1, Lakeside alder community
 68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	0-30	76	
	30-35	44	
	40-50	45	
	50-60	38	
	60-70	51	
	70-80	83	25
	80-90	65	19
	90-100	72	19
	100-110	75	19
	110-120	72	17
	120-130	48	0
	130-140	47	0
	140-150	46	0
	150-160	43	0
	160-170	44	0
	170-180	41	0
	180-190	45	0
	190-200	54	0

C-la2, Lakeside alder community
 68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	26-40	58	
	40-50	45	
	50-60	114	55
	60-70	131	46
	70-80	70	14
	80-90	136	47
	90-100	114	28
	100-110	98	21
	110-120	62	17
	120-130	53	8
	130-140	82	49
	140-150	68	17
	150-160	44	0
	160-170	35	0
	170-180	33	0
	180-190	32	0

C-la3, Lakeside alder community
 68°43'00"N, 134°16'30"W

Aug-99	Depth (cm)	MC	EI
	0-20	57	
	20-30	61	
	30-40	71	
	40-50	160	21
	50-60	198	27
	60-70	174	35
	70-80	169	42
	80-90	109	34
	90-100	105	35
	100-110	139	40
	110-120	140	40
	120-130	116	33
	130-140	130	38
	140-150	87	23
	150-160	158	34
	160-170	55	9
	170-180	60	17
	180-190	84	8
	190-200	82	10
	200-210	69	0
	210-220	67	5
	220-230	76	14
	230-240	73	0
	240-250	57	3

C-sab1, Spruce/alder-bearberry community
68°43'00"N, 134°16'30"W

Aug-99	Depth (cm)	MC	EI
	0-24	61	
	23-40	46	
	40-60	32	
	60-80	38	
	80-90	23	
	90-100	59	7
	100-110	53	10
	110-120	65	20
	120-130	54	15
	130-140	50	5
	140-150	46	4
	150-160	42	2
	160-170	36	0
	170-180	35	0
	180-190	37	0
	190-200	33	0
	200-210	41	0

C-sab2, Spruce/alder-bearberry community
68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	0-35	67	0
	35-50	58	0
	50-60	33	0
	60-70	30	0
	70-80	25	0
	80-90	51	0
	90-100	116	16
	100-110	88	11
	110-120	56	5
	120-130	52	8
	130-140	58	7
	140-150	55	5

C-sf1, Spruce/feathermoss community
68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	23-30	86	
	30-40	44	
	40-50	41	
	50-60	69	30
	60-70	103	32
	70-80	83	27
	80-90	97	33
	90-100	97	38
	100-110	98	36
	110-120	78	23
	120-130	98	35
	130-140	85	31
	140-150	84	23
	150-160	54	10
	160-170	58	12
	170-180	63	15
	180-190	62	11
	190-200	58	2

C-sf2, Spruce/feathermoss community
68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	30-35	101	0
	40-50	21	0
	50-60	43	0
	60-70	103	42
	70-80	90	29
	80-90	95	35
	90-100	130	42
	100-110	135	37
	110-120	91	31
	120-130	99	36
	130-140	77	21
	140-150	93	28
	150-160	63	17
	160-170	48	0

C-sf3, Spruce/feathermoss community

68°43'00"N, 134°16'30"W

Jul-99	Depth (cm)	MC	EI
	0-30		
	30-40	77	
	40-50	33	
	50-60	37	
	60-70	81	19
	70-80	151	25
	80-90	146	41
	90-100	150	31
	100-110	95	32
	110-120	124	29
	120-130	76	27
	130-140	65	20
	140-150	80	22
	150-160	93	33
	160-170	79	21
	170-180	48	0
	180-190	65	19
	190-200	54	34

C-sf4, Spruce/feathermoss community

68°43'00"N, 134°16'30"W

Aug-99	Depth (cm)	MC	EI
	0-20	49	
	20-40	61	
	40-45	50	
	45-50	35	
	50-60	74	12
	60-70	92	31
	70-80	81	27
	80-90	55	13
	90-100	62	14
	100-110	96	40
	110-120	74	25
	120-130	73	34
	130-140	87	32
	140-150	66	16
	150-160	54	17
	160-170	45	3
	170-180	48	2
	180-190	45	4
	190-200	46	0

C-sf5, Spruce/feathermoss community

68°43'00"N, 134°16'30"W

Aug-99	Depth (cm)	MC	EI
	0-20	62	
	20-40	66	
	40-60	32	
	60-65	36	
	65-70	55	14
	70-80	110	50
	80-90	88	35
	90-100	67	27
	100-110	81	24
	110-120	73	31
	120-130	67	17
	130-140	55	5
	140-150	52	8
	150-160	54	5
	160-170	45	0
	170-180	43	0
	180-190	40	0
	190-200	44	0

C-sf6, Spruce/feathermoss community

68°43'00"N, 134°16'30"W

Aug-99	Depth (cm)	MC	EI
	0-20	154	
	20-40	26	
	40-45	42	
	45-50	93	32
	50-60	106	38
	60-70	177	53
	70-80	81	27
	80-90	61	14
	90-100	61	12
	100-110	67	23
	110-120	67	15
	120-130	93	25
	130-140	52	8
	140-150	44	5
	150-160	40	1
	160-170	44	1

Central Delta
(Inuvik region) (see Figure 3.1)

C-w7, Point-bar willow community
 68°24'00"N, 133°47'40"W

Apr-00	Depth (cm)	MC	EI
	0-10	79	
	10-20	66	
	20-30	47	
	30-40	41	
	40-50	42	
	50-60	31	
	60-70	36	
	70-80	22	
	80-90	15	
	90-100	22	
	100-110	19	
	110-120	29	
	120-130	41	0

C-sab3, Spruce/alder-bearberry community
 68°24'30"N, 133°47'50"W

Jul-99	Depth (cm)	MC	EI
	40-50	21	
	50-60	24	
	60-70	14	
	70-80	22	
	80-90	30	
	90-100	12	
	100-110	25	
	110-120	21	
	120-130	27	0
	130-140	39	0
	140-150	41	0
	150-160	38	0
	160-170	45	0
	170-180	47	0
	180-190	37	0
	190-200	39	0
	200-210	31	0
	210-220	33	0
	220-230	34	0
	230-240	38	0

C-sab4, Spruce/alder-bearberry community
 68°24'30"N, 133°47'50"W

Jul-00	Depth (cm)	MC	EI
	0-20	49	
	20-40	45	
	40-50	42	
	50-60	31	
	60-70	21	
	70-80	22	
	80-90	26	
	90-100	22	
	100-110	37	0
	110-120	42	0
	120-130	44	5
	130-140	43	0
	140-150	40	0

C-sf7, Spruce/feathermoss community
68°22'40"N, 133°46'30"W

Jul-99	Depth (cm)	MC	EI
	30-50	82	**
	50-70	78	**
	70-90	76	**
	90-110	50	**
	110-130	37	**
	130-150	44	**
	150-170	47	**
	170-190	38	**
	190-210	42	**
	210-230	51	**
	230-250	43	**

C-sf8, Spruce/feathermoss community
68°22'40"N, 133°46'30"W

Jul-00	Depth (cm)	MC	EI
	0-20	46	
	20-40	64	
	40-50	68	4
	50-60	158	47
	60-70	140	41
	70-80	133	41
	80-90	167	47
	90-100	118	41
	100-110	178	53
	110-120	95	30

** Excess ice content was not determined for these samples

C-sc1, Spruce/crowberry-lichen community
68°26'30"N, 133°50'00"W

Jul-99	Depth (cm)	MC	EI
	30-40	36	
	40-50	141	49
	50-60	122	47
	60-70	96	41
	70-80	110	33
	80-90	69	27
	90-100	105	43
	100-110	102	36
	110-120	63	20
	120-130	87	28
	130-140	59	21
	140-150	94	38
	150-160	49	10
	160-170	70	22
	170-180	42	8
	180-190	62	32
	190-200	82	29
	200-210	45	8
	210-220	50	0
	220-230	41	0
	230-240	57	0
	240-250	36	0
	250-260	36	0
	260-270	43	0
	270-280	35	0
	280-290	43	0
	290-300	36	0
	300-310	33	0
	310-320	32	0
	320-330	39	0

C-sc2, Spruce/crowberry-lichen community
68°26'30"N, 133°50'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	47	
	20-40	24	
	40-50	25	
	50-60	28	
	60-70	31	
	70-80	79	19
	80-90	91	33
	90-100	109	41
	100-110	206	59
	110-120	180	53

C-sc3, Spruce/crowberry-lichen community
68°26'30"N, 133°50'00"W

Apr-01	Depth (cm)	MC	EI
	20-40	126	0
	50-60	128	42
	60-70	208	55
	70-80	100	33
	80-90	96	30
	90-100	107	44

C-sc4, Spruce/crowberry-lichen community
68°26'30"N, 133°50'00"W

Aug-99	Depth (cm)	MC	EI
	0-20	230	
	20-40	47	
	40-50	57	
	50-60	110	35
	60-70	118	45
	70-80	141	50
	80-90	114	38
	90-100	110	36
	100-110	121	38
	110-120	106	25
	120-130	74	6
	130-140	80	18
	140-150	63	5
	150-160	46	0

C-sc5, Spruce/crowberry-lichen community
68°27'00"N, 133°51'00"W

Aug-00	Depth (cm)	MC	EI
	0-50	70	
	50-60	54	5
	60-70	149	47
	70-80	119	40
	80-90	154	50
	90-100	181	54
	100-110	179	52
	110-120	73	13
	120-130	52	3

C-sc6, Spruce/crowberry-lichen community
68°26'00"N, 133°48'30"W

Aug-99	Depth (cm)	MC	EI
	0-30	261	
	30-40	399	
	40-50	270	16
	50-60	155	52
	60-70	133	45
	70-74	280	57
	80-90	137	41
	90-100	108	39
	100-110	97	30
	110-120	77	18
	120-130	63	12
	130-140	79	23
	140-150	65	10

C-Ia4, Lakeside alder community
 68°22'20"N, 133°46'00"W

Jul-99	Depth (cm)	MC	EI
	40-50	56	
	50-60	262	34
	60-70	295	26
	70-80	132	44
	80-90	186	55
	90-100	190	22
	100-110	278	70
	110-120	141	35
	120-130	179	25
	130-140	273	29
	140-150	255	27
	150-160	297	36
	160-170	304	50
	170-180	269	47
	180-190	202	43
	190-200	211	48
	200-210	208	45
	210-220	174	30
	220-230	194	18
	230-240	253	30
	240-250	216	42
	250-260	148	27
	260-270	136	30
	270-280	105	31
	280-290	72	21

1c. Outer (Northern) Delta (see Figure 4.3)

Key	MC = Gravimetric moisture content (%)
	EI = Excess ice content (%)
	Bold = permafrost

O-w1, Point-bar willow community
69°10'30"N, 135°27'30"W

Jul-00	Depth (cm)	MC	EI
	0-20	32	
	20-40	33	
	40-60	33	
	60-70	35	
	70-80	36	
	80-90	33	
	90-100	33	
	100-110	30	0
	110-120	30	0
	120-130	37	0
	130-140	32	0
	140-150	33	0
	150-160	32	0
	160-170	31	0
	170-180	39	0

O-w2, Point-bar willow community
69°10'30"N, 135°27'30"W

Jul-00	Depth (cm)	MC	EI
	0-20	59	
	20-40	43	
	40-50	44	
	50-60	37	
	60-70	37	
	70-80	34	
	80-90	30	
	90-100	22	
	100-110	19	
	110-120	27	0
	120-130	32	0
	130-140	**	0
	140-150	**	0
	150-160	33	0
	160-170	32	0
	170-180	32	0

O-w3, Point-bar willow community
69°10'30"N, 135°27'30"W

Jul-00	Depth (cm)	MC	EI
	0-20	63	
	20-40	41	
	40-50	38	
	50-60	40	
	60-70	37	
	70-80	36	
	80-90	36	
	90-100	33	
	100-110	38	0
	110-120	38	0
	120-130	39	0
	130-140	36	0
	140-150	34	0
	150-160	39	0
	160-170	36	0
	170-180	**	0
	180-190	34	0

O-w4, Point-bar willow community
69°10'30"N, 135°27'30"W

Jul-00	Depth (cm)	MC	EI
	0-20	38	
	20-40	36	
	40-50	35	
	50-60	42	
	60-70	36	
	70-80	47	
	80-90	34	
	90-100	32	0
	100-110	36	0
	110-120	37	0
	120-130	41	0
	130-140	31	0
	140-150	35	0
	150-160	34	0
	160-170	31	0
	170-180	35	0
	180-190	35	0
	190-200	33	0

O-w5, Point-bar willow community

69°10'30"N, 135°26'00"W

Jul-00	Depth (cm)	MC	EI
	0-20	47	
	20-40	36	
	40-50	37	
	50-60	35	
	60-70	36	
	70-80	33	
	80-90	33	
	90-100	34	
	100-110	32	0
	110-120	28	0
	120-130	24	0
	130-140	32	0
	140-150	32	0
	150-160	37	0
	160-170	38	0
	170-180	32	0
	180-190	29	0
	190-200	36	0

O-s1, Sedge community

69°10'30"N, 135°26'00"W

Aug-01	Depth (cm)	MC	EI
	0-20	48	
	20-40	42	
	40-60	44	
	60-70	45	3
	70-80	48	5
	80-90	51	10
	90-100	137	48
	100-110	125	46
	110-120	105	42
	120-130	138	47
	130-140	170	52
	140-150	108	33

O-s2, Sedge community

69°08'40"N, 135°27'30"W

Jul-00	Depth (cm)	MC	EI
	0-20	88	
	20-30	58	
	30-40	67	
	40-50	104	22
	50-60	128	29
	60-70	74	11
	70-80	81	18
	80-90	143	25
	90-100	185	27
	100-110	289	48
	110-120	240	43
	120-130	161	43
	130-140	157	52
	140-150	74	35
	150-160	57	18
	160-170	60	25
	170-180	64	33
	180-190	98	47
	190-200	106	43
	200-210	63	21
	210-220	65	27
	220-230	46	12
	230-240	47	15

APPENDIX 2. ACTIVE-LAYER AND PERMAFROST MOISTURE AND ICE CONTENT, HUMMOCKY TERRAIN, NAVY ROAD, INUVIK

2a. Hummock tops

Key	MC = Gravimetric moisture content (%)
	EI = Excess ice content (%)
	Bold = permafrost

N1, Navy Road, hummock top

Jul-99	Depth (cm)	MC	EI
	0-20	30	
	20-40	32	
	40-50	34	
	50-60	34	
	60-70	35	
	70-90	37	
	90-100	56	
	100-110	108	8
	110-120	66	1
	120-130	57	3
	130-140	139	12
	140-150	175	36
	150-160	117	29
	160-170	48	6
	170-180	61	4
	180-190	115	8
	190-200	134	13
	200-210	159	6
	210-220	68	3
	220-230	33	0

N2, Navy Road, hummock top

Aug-00	Depth (cm)	MC	EI
	0-20	35	
	20-40	35	
	40-60	34	
	60-70	29	
	70-80	31	
	80-90	27	
	90-100	44	
	100-110	120	20
	110-120	52	0
	120-130	50	0
	130-140	156	35
	140-150	149	34
	150-160	139	30
	160-170	90	24
	170-180	65	4
	180-190	106	9
	190-200	246	28
	200-210	183	13

N3, Navy Road, hummock top

Aug-00	Depth (cm)	MC	EI
	0-20	34	
	20-30	38	
	30-40	40	
	40-50	35	
	50-60	31	
	60-70	29	
	70-80	27	
	80-85	29	
	85-90	41	
	90-100	58	1
	100-110	115	25
	110-120	104	18
	120-130	76	5
	130-140	112	18
	140-150	119	32
	150-160	86	25
	160-170	76	21
	170-180	90	21
	180-190	95	23
	190-200	102	19
	200-210	117	19
	210-220	149	25
	220-230	43	0

N4, Navy Road, hummock top

Jul-00	Depth (cm)	MC	EI
	0-20	34	
	20-40	35	
	40-60	31	
	60-70	33	
	70-80	38	
	80-90	41	
	90-100	58	3
	100-110	138	33
	110-120	64	0
	120-130	71	6
	130-140	170	43
	140-150	185	45

2b. Interhummock troughs**N5, Navy Road, interhummock trough**

Jul-99	Depth (cm)	MC	EI
	0-26	418	
	20-40	132	
	40-50	231	3
	50-60	105	1
	60-70	73	2
	70-80	63	2
	80-90	83	7
	90-100	134	25
	100-110	55	8
	110-120	80	17
	120-130	135	34
	130-140	160	34
	140-150	210	44
	150-160	106	17
	160-170	135	27
	170-180	162	11
	180-190	131	4

N6, Navy Road, interhummock trough

Jul-99	Depth (cm)	MC	EI
	0-28	247	
	30-40	471	
	40-50	92	
	50-60	273	0
	60-70	312	0
	70-80	148	0
	80-90	80	3
	90-100	108	19
	100-110	87	7
	110-120	283	40
	120-130	138	35
	130-140	89	22
	140-150	93	9
	150-160	84	5
	160-170	142	14
	170-180	102	0
	180-190	120	5
	190-200	15	0

N7, Navy Road, interhummock trough

Aug-01	Depth (cm)	MC	EI
	20-30	400	
	30-40	335	
	40-50	213	6
	50-60	193	34
	60-70	216	8
	70-80	223	10
	80-90	145	35
	90-100	252	23
	100-110	238	24
	110-120	121	32
	120-130	88	25
	130-140	80	15
	140-150	55	0
	150-160	95	20
	160-170	103	10

2c. Degrading permafrost experiment

Moisture content profiles beneath degrading permafrost following removal of hummock top, N2

N8, Navy Road, hummock top

Sept-01	Depth (cm)	MC
	115-120	63
	120-125	42
	125-130	49
	130-140	137

N9, Navy Road, hummock top

Sept-01	Depth (cm)	MC
	115-120	59
	120-125	45
	125-130	46
	130-135	85

2d. Aggrading permafrost experiment

Core N4 was obtained from a hummock top early in summer 2000.

The hole (N2) was backfilled and organic materials were added to the hummock top causing the permafrost table to aggrade. Core N4b was obtained in September of 2001.

Core N4		Core N4b	
Jul-00		Sept-01	
Depth (cm)	MC	Depth (cm)	MC
0-20	34	0-20	32
20-40	35	20-40	33
40-60	31	40-60	35
60-70	33	60-80	31
70-80	38		
80-90	41	80-90	39
90-100	58	90-100	60
100-110	138	100-110	154
110-120	64		
120-130	71		
130-140	170		
140-150	185		

APPENDIX 3. GEOCHEMISTRY OF ACTIVE-LAYER AND PERMAFROST SEDIMENTS.

3a. Mean total soluble cation concentrations, active layer and permafrost, and enrichment ratios.

Sites with medium to high ice-content permafrost (>10% excess ice content) are indicated in bold.

Core ID	Mean total soluble cation concentration (meq 100g ⁻¹ dry soil)		Enrichment ratio N-S permafrost/AL
	Active layer	Near-surface permafrost	
S-sf1	0.31	1.19	3.8
S-sab1	0.38	0.38	1.0
S-pl	0.29	0.61	2.1
S-a1	0.41	0.53	1.3
S-sf3	0.32	0.83	2.6
C-w1	1.64	1.12	0.7
C-w2	2.06	3.83	1.9
C-w6	1.39	3.07	2.2
C-a2	1.54	2.15	1.4
C-sab1	0.64	2.09	3.3
C-sab4	2.03	4.93	2.4
C-sf4	0.49	1.24	2.5
C-sc2	0.21	0.58	2.8
C-sf8	0.32	0.81	2.5
C-sc5	0.28	2.25	8.1
C-sc8	0.49	1.59	3.3
C-sc7	0.24	1.70	7.0
C-sc9	0.50	1.38	2.8
C-sc10	0.37	1.62	4.3
C-sc11	0.52	1.64	3.1
O-s2	0.57	1.42	2.5
O-w5	0.42	0.47	1.1
O-Pin1	0.39	1.95	5.0
O-Pin2	0.65	0.77	1.2
N2	0.14	0.40	2.9
N1	0.11	0.17	1.5
N3	0.04	0.12	2.7
N4	0.04	0.09	2.3
DH-1	0.03	0.09	3.1
IS-1	0.29	0.66	2.3

* Enrichment ratios for each core were calculated by dividing mean total soluble cation concentration in the top 30 to 50 cm of permafrost by the mean concentration in overlying mineral soils of the active layer.

** C-sc7 to C-sc11 are from a spruce/crowberry lichen forest on East Channel, 15 km north of Inuvik. The site is indicated 'D1' on Figure 3.1.

DH - 1 is upland terrain within the Delta

68°24'00"N, 133°49'00"W

IS-1 is a black spruce forest site 3 km south of Inuvik.

68°19'30"N, 133°39'00"W

3b. Post-fire release of soluble nutrients and water from near-surface permafrost

A) Hummocky terrain, black spruce woodland, Navy Road, Inuvik (Site N).

1A. Amounts of soluble cations contributed to the base of the active layer following degradation of near-surface permafrost, Cores N1, N3 and N4, hummock tops, Navy Road, Inuvik.

Year	AL deepening (cm)	$\text{Ca}^{++} \text{ g m}^{-2}$			Ave	$\text{Mg}^{++} \text{ g m}^{-2}$			Ave	$\text{Na}^{+} \text{ g m}^{-2}$			Ave	$\text{K}^{+} \text{ g m}^{-2}$			Ave
		N1	N3	N4		N1	N3	N4		N1	N3	N4		N1	N3	N4	
1 to 5	57	4.39	5.39	3.03	4.27	1.23	0.46	0.42	0.70	14.02	5.67	5.78	8.49	16.30	15.34	8.49	13.38
1	25	1.93	2.37	1.33	1.87	0.54	0.20	0.18	0.31	6.15	2.49	2.54	3.72	7.15	6.73	3.72	5.87
2	12	0.92	1.14	0.64	0.90	0.26	0.10	0.09	0.15	2.95	1.19	1.22	1.79	3.43	3.23	1.79	2.82
3	8	0.62	0.76	0.43	0.60	0.17	0.06	0.06	0.10	1.97	0.80	0.81	1.19	2.29	2.15	1.19	1.88
4	8	0.62	0.76	0.43	0.60	0.17	0.06	0.06	0.10	1.97	0.80	0.81	1.19	2.29	2.15	1.19	1.88
5	4	0.31	0.38	0.21	0.30	0.09	0.03	0.03	0.05	0.98	0.40	0.41	0.60	1.14	1.08	0.60	0.94

**Estimates based on mean cation concentrations of core samples obtained from the top 50 cm of permafrost.

2A. Mean ice content in top 50 cm of permafrost, Cores N1, N3 and N4, hummock tops, Navy Road, Inuvik.

Core	Mean ice content (%)
N1	12
N3	19
N4	26
Ave.	19

3A. Amounts of moisture contributed to the base of the active layer following degradation of near-surface permafrost in hummocky terrain.

Year	AL deepening (cm)	Moisture released from permafrost (cm)
1 to 5	57	12.0
1	25	5.3
2	12	2.5
3	8	1.7
4	8	1.7
5	4	0.8

**Calculations using mean excess ice content estimated in Table 2A.

B) White spruce/crowberry-lichen forest, East Channel, Mackenzie Delta (Site D1)

1B. Amounts of soluble cations contributed to the base of the active layer following degradation of near-surface permafrost, Cores C-sc7, C-sc8, C-sc9, spruce/crowberry-lichen forest, Site D1, East Channel, Mackenzie Delta.

Year Sum	AL deepening (cm)	Ca ⁺⁺ g m ⁻²		Ave	Mg ⁺⁺ g m ⁻²		Ave	Na ⁺ g m ⁻²		Ave	K ⁺ g m ⁻²		Ave
		C-sc7	C-sc8	C-sc9	C-sc7	C-sc8	C-sc9	C-sc7	C-sc8	C-sc9	C-sc7	C-sc8	C-sc9
1 to 5	57	111.788	101.06	82.3	76.84	74.17	64.66	9.76	11.45	10.21	10.00	4.77	12.80
1	25	49.03	44.32	36.11	33.70	32.53	28.36	4.28	5.02	4.48	4.38	2.09	5.61
2	12	23.53	21.28	17.34	16.18	15.62	13.61	2.05	2.41	2.15	2.10	1.00	2.69
3	8	15.69	14.18	11.56	10.79	10.41	9.07	1.37	1.61	1.43	1.40	0.67	1.80
4	8	15.69	14.18	11.56	10.79	10.41	9.07	1.37	1.61	1.43	1.40	0.67	1.80
5	4	7.84	7.09	5.78	5.39	5.21	4.54	0.68	0.80	0.72	0.70	0.33	0.90
				6.90			5.04			0.73			0.64

**Estimates based on mean cation concentrations of core samples obtained from the top 30 to 50 cm of permafrost.

2B. Mean ice content in top 30 to 50 cm of permafrost, Cores C-sc7, C-sc8 and C-sc9, site D1, spruce/crowberry-lichen forest, Mackenzie Delta.

Core	Mean ice content (%)
C-sc7	45
C-sc8	32
C-sc9	39
Ave.	39

3B. Amounts of moisture contributed to the base of the active layer following degradation of near-surface permafrost, Cores C-sc7, C-sc8 and C-sc9, site D1, spruce/crowberry-lichen forest, Mackenzie Delta.

Year	AL deepening (cm)	Moisture released from permafrost (cm)
1 to 5	57	32.8
1	25	14.4
2	12	6.9
3	8	4.6
4	8	4.6
5	4	2.3

**Calculations using mean excess ice content estimated in Table 2B.

APPENDIX 4. NEAR-SURFACE GROUND ICE AND SPRUCE FOREST STRUCTURE, MACKENZIE DELTA.

Ice content of near-surface permafrost and mean tree tilt, Mackenzie Delta (see Figure 4.5).

Core ID	Mean ice content in top 50 cm of permafrost (%)	Mean site tree tilt
S-sab1	3	2
S-sf1	31	9
S-sf3	15	4
S-sf4	26	3
C-sab1	11	1
C-sab2	9	2
C-sf1	32	9
C-sf3	29	6
C-sf4	19	5
C-sf5	33	7
C-sf6	33	9
C-sab3	0	1
C-sab4	1	0
C-sf8	43	12
C-sc1	39	15
C-sc3	41	16
C-sc5	49	15
C-sc6	42	11