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ON THE ORIGIN OF AGGRADATIONAL ICE IN PERMAFROST

by

CHRISTOPHER ROBERT BURN, B.Sc., M.A.

A thesis submitted to
the Faculty of Graduate Studies and Research,
Carleton University,
in partial fulfilment of
the requirements for the degree of

Doctor of Philosophy.

Department of Geology,
Carleton University,
Ottawa, Ontario

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

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submitted by

CHRISTOPHER ROBERT BURN, B.Sc., M.A.

in partial fulfillment of the requirements
for the degree of Doctor of Philosophy.

Chairman, Department of Geology

Thesis Supervisor

External Examiner

Carleton University

12 September 1986

ABSTRACT

An investigation of the annual water balance of the near surface layers was conducted at eight sites near Mayo, Yukon Territory, to ascertain the origin of the ice-rich zone commonly found beneath the active layer in permafrost terrain.

Water movement in frozen ground is principally driven by temperature-induced potential gradients, which move water downward into permafrost in summer and fall, and upward out of permafrost in winter. Less winter desiccation than summer infiltration occurs, due to the lower hydraulic conductivity at the prevailing temperatures. Overall, permafrost is a sink for atmospheric water.

Direct measurements of the thermal and hydrological regimes at the sites were accompanied by measurements of soil heaving, which indicate water accumulation in particular soil layers. The hydrological processes were simulated on a short-term basis in PVC tubes filled with soil slurry. In addition, the tritium content of permafrost and the ground ice stratigraphy in the study area were examined for evidence of long-term moisture movement into permafrost.

The results indicate that downward water movement into frozen ground occurs throughout the summer. During freeze-back, however, most water movement is upward, and results in frost heave. Up to 23 percent of the frost heave measured occurred within frozen

soil, independent of the penetration of the frost front. Throughout the winter, upward water movement continued in the ground, driven by the soil temperature gradient. Overall, a mean maximum rate of water incorporation into permafrost of 3.5 mm year^{-1} was detected.

Tritium, derived from nuclear testing in the 1950s and 1960s, has infiltrated permafrost to depths of up to 50 cm. The distribution of tritium in permafrost at two sites irrigated with tritiated water in 1983 suggests that tritium infiltration is temperature-induced. Observations of the ground ice stratigraphy in the headwall of a thaw slump indicate that the processes considered have been continuing throughout the Holocene. The 90 cm of segregated ice above a thaw unconformity, dated at 8870 ± 200 years BP, provides a mean minimum rate of permafrost aggradation of between 0.1 and 0.2 mm year^{-1} over the past 8 ka.

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Appendix

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during soil freezing. A 3

NOMENCLATURE

C_a	Apparent heat capacity.
C_m	Volumetric heat capacity of soil-water-ice mixture
$D(w)$	Soil-water diffusivity.
$D(w)_v$	Diffusivity of the vapour phase.
$D(r)$	Diffusivity of tritium in soil-water-ice system.
e	Saturation vapour pressure over ice.
K	Hydraulic conductivity of unfrozen soil.
K_f	Hydraulic conductivity of frozen soil.
k	Thermal conductivity.
L	Latent heat of fusion of water.
P_i	Ice pressure.
P_w	Unfrozen water pressure.
ψ	Soil water potential.
Q	Flow rate.
Q_l	Flow rate of liquid phase.
Q_v	Flow rate of vapour phase.
$Q(r)$	Flow of tritium.
R	Gas constant for water vapour.
r	Ratio of observed to standard counts.
ρ_w	Dry bulk density.
T	Temperature.
t	Time.
r	Tritium concentration.
θ_u	Unfrozen water content (volumetric).
w	Gravimetric water content.

w_u Unfrozen water content (gravimetric).
 v_i Specific volume of ice.
 v_w Specific volume of water.
 z Distance.

Frontispiece. Oblique aerial photograph of the larger
thaw slump in the Mayo study area, with
thermokarst lake behind. 12 August 1984.

CHAPTER 1

INTRODUCTION

The purpose of this thesis is to examine the mechanisms determining the water balance of permafrost in order to explain the development of near-surface ground ice. The operation of these processes throughout the Holocene was examined by field experiments and observations of ground ice stratigraphy at a site near Mayo, central Yukon Territory (Figure 1.1). The processes examined here are ubiquitous in perennially frozen ground, and since one fifth of the world's land surface (including half of Canada) is underlain by permafrost (R.J.E. Brown 1970), they occur over a substantial portion of the earth's surface.

Permafrost is particularly significant to hydrology, for at temperatures below 0°C thermal gradients control water movement in the ground (Parmuzina 1978; Mackay 1983a). In comparison, hydrostatic forces are frequently negligible at these temperatures (Harlan 1974; Williams 1982).

1.1 Permafrost hydrology.

This thesis is concerned with the movement of water in the surface layers of permafrost terrain. The processes examined are not adequately treated in the established literature concerned

Figure 1.1 Permafrost map of the Yukon Territory.
(after Brown 1978).

with the groundwater or watershed hydrology of permafrost regions, where the frost table is simply viewed as an impermeable barrier, and permafrost as an aquiclude (e.g. Brandon 1965; Williams 1970; Rouse 1982a; Slaughter et al. 1983; Woo and Steer 1983). The present study draws on the literature concerned with frost heaving and the movement of water in freezing and frozen soils (e.g. Burt and Williams 1976; Miller 1978; Horiguchi and Miller 1980; Mackay 1983a).

1.1.1 Water movement in freezing soils.

It has been known for some time that not all water in soil freezes at 0°C (Nersesova 1953). Moreover, the relationship between water and ice in frozen soil is similar to that between water and air in unfrozen soil (Koopmans and Miller 1966), and results in a suction gradient in the unfrozen water towards regions of lower temperature (Williams 1982). The suction gradient is directly proportional to the temperature gradient, with a constant of proportionality of approximately $1.2 \text{ MPa } ^{\circ}\text{C}^{-1}$, or $120 \text{ m(H}_2\text{O) } ^{\circ}\text{C}^{-1}$ (Edelfsen and Anderson 1943).

The existence of unfrozen water in soils below 0°C also suggests that these materials may be permeable. Williams and Burt (1974; Burt and Williams 1976) were the first to measure the hydraulic conductivity of soils at temperatures below 0°C , and recorded a steep decline in the permeability of soil with temperature below 0°C . However, they were still able to determine values of about $10^{-10} \text{ m s}^{-1}$ below -0.3°C .

4

The coupling of suction gradients with hydraulic conductivity results in the processes leading to ice segregation and the development of ice lenses in frozen soil (Mackay 1972; Miller 1978).

1.2 Ground ice.

Permafrost derives its principal physical significance from the presence of ice in soils at volumes greater than the soil pore space (French 1976; Williams 1982). Ice present in frozen soil at a volume greater than soil saturation is referred to as excess ice. Ice masses found below the ground surface are referred to as ground ice.

There are several geomorphological processes that lead to the formation of ground ice (Mackay 1972): burial, ~~and~~ ice-cored moraine (Sugden and John 1976), buried snowbanks (French and Pollard 1986); ice wedging (Mackay 1974a, 1984; Mackay and Matthews 1983); ice intrusion, e.g. some pingo ice (Mackay 1973, 1979), frost blisters (van Everdingen 1978, 1982; Pollard and French 1983, 1984); and ice segregation, e.g. reticulated vein ice (Mackay 1974b), massive ice (Mackay 1971a). In addition, Brown et al. (1978) describe the form "aggradational" ice, which they consider to be presently forming where a rising permafrost table traps moisture at the base of the annually-thawed active layer, incorporating the water into permafrost (Mackay 1971b, 1972, 1973b).

1.2.1 Aggradational ice.

Aggradational ice is almost ubiquitous within permafrost terrain, in the form of an ice-rich zone at the base of the active layer (the surface layer which thaws each summer; see Figure 1.2 and Mackay 1975; Pollard and French 1980; Outcalt 1982). Recently, Cheng (1982, 1983) has suggested that the accumulation of ice in this zone may be part of the annual water balance of permafrost terrain, rather than simply a product of a cooling climate or a period of rising surface level.

At the base of the active layer, most water movement is controlled by temperature-induced suction gradients, rather than gravity (Cheng 1983; Mackay 1983a). In summer, these gradients tend to draw water downward into frozen ground ahead of the descending thawing front (Mackay 1983a). In winter, the temperature gradients induce water flow in the opposite direction (Parmuzina 1978; Mackay *et al.* 1979), but the low hydraulic conductivity at the prevailing winter temperatures (Burt and Williams 1976; Horiguchi and Miller 1983; Smith 1985a) results in a net imbalance between summer and winter fluxes, and the permafrost zone appears as a net sink for active layer water (Cheng 1983).

The hydrological regime of permafrost is, therefore, intimately linked to the ground thermal regime (Wright 1981), and varies as the conditions controlling the thermal regime change.

Figure 1.2 Gravimetric water content of two 5 m drill cores from Mayo study area, 1985.

1.3 Permafrost distribution.

The ground thermal regime at a site is a response to the surface temperature regime, the thermal properties of earth materials at the site and the flow of heat from the earth's interior (Lachenbruch et al. 1962; Washburn 1979). The critical variable in terms of permafrost distribution is the ground surface temperature, which is a function of the surface energy balance (Geiger 1965; Oke 1978). Hence, the distribution and thickness of permafrost is not only dependent on the regional climate, but also on site-specific, microclimatic factors, particularly snow depth (Smith 1975, 1976; Goodrich 1982) and surface wetness (Smith and Riseborough 1983).

Permafrost terrain is characterized by a surface active layer, that thaws each summer, underlain by perennially frozen ground. Active layer thickness varies within a small area (Owens and Harper 1977), depending primarily on the summer surface temperature (Mackay 1970), the thickness of the soil surface organic matter (R.J.E. Brown 1970; Zoltai and Tarnocai 1975; Riseborough 1985) and the water (or ice) content of soil materials (Farouki 1981). Ferrians and Hobson (1973) have suggested a global range of 15 cm to 5 m for active layer thickness.

Permafrost thickness is measured from the base of the active layer, at the permafrost table, to the point where the temperature again rises above 0°C (Figure 1.3). The thickness varies from a few centimetres at the margins of permafrost, to over 1400 m in

Figure 1.3 Generalised thermal regime of permafrost.

(R.J.E. Brown 1970, Fig. 6).

Siberia (Grave 1968). Thicknesses of over 700 m have been reported in Arctic Canada by Taylor et al. (1982). In regions of shallow permafrost, thickness depends upon the present surface temperature regime; larger thicknesses may also reflect past surface climates (Cermak 1971; Lachenbruch et al. 1982; Judge and Taylor 1985).

The permafrost map of Canada (Brown 1978) demarcates four zones of permafrost--continuous, widespread, scattered and alpine--on the basis of mean annual air temperature. The borders of the widespread permafrost zone, in which Mayo is located (see Figure 1.1), are drawn along the -5°C and -3.9°C isotherms. Since the ground temperature is not related to air temperature in a straight-forward fashion (Judge 1973; Smith 1975, 1976), sudden changes in ground thermal regime do not occur at the borders of these zones (Williams 1979). Instead, the zones represent the transition from the seasonally frozen ground of most of southern Canada to the perennially frozen soils of the Arctic.

1.4 Study area.

Ground temperatures measured at 5 m depth in undisturbed terrain at six sites in the study area between June 1983 and September 1984 have indicated a range in mean annual temperature of -0.8°C to -2.2°C (mean -1.6°C) (Smith and Burn 1986). Combining the mean value with a geothermal gradient of $43\text{ m }^{\circ}\text{C}^{-1}$ for sites in southern and central Yukon (Burgess et al. 1982), the current equilibrium permafrost depth in the area is approximately 70 m.

Most of the records of permafrost thickness in the Mayo area are from wells drilled for residents. Ground ice has been reported to depths of over 130 ft (40 m) in a well drilled at the Mayo Group Home (see Figure 1.4), located in similar terrain to that of the study area. Other residents have encountered groundwater three or four metres below the surface.

The precise study area is located in a body of ice-rich glaciolacustrine sediments (Hughes 1983a), which contains several thermokarst lakes and two principal exposures of ground ice in the vertical headwalls of two thaw slumps (see Figure 1.4). The larger of these is over 100 m long and 6 m high (see Frontispiece). Segregated ice of various cryotextures and truncated ice wedges are exposed in this wall along with several organic units that have been interpreted as thermokarst lake bottoms and ^{14}C dated at about 8.5 ka BP (French *et al.* 1984; Burn *et al.* 1986).

The water content of much of the icy sediment in the study area is several times greater than saturation, up to 90 percent ice by volume (Smith *et al.* 1983).

1.5 Field experiments.

This study documents the annual mass balance of the surface 2 m of eight sites near Mayo, in order to examine Cheng's hypothesis, as outlined above. Seven of the sites have a permafrost regime, and, of these, six are in glaciolacustrine, frost susceptible sediments. The seventh is in the fine sand and silt of the Stewart River flood plain. The other site is located in the

Figure 1.4. Aerial photo mosaic of the Mayo area.

sandy sediments of a recently (1952) drained lake (Burn 1982), and is only frozen seasonally.

In addition, evidence concerning moisture movement into permafrost over the Holocene period has been obtained from the ground ice stratigraphy exposed in the headwall of the larger thaw slump. Investigations of the infiltration of tritium (^3H) into permafrost since the atmospheric atomic bomb tests of the late 1950s and early 1960s (following Michel and Fritz 1978; Chizhov et al. 1983) have provided information regarding water movement at the base of the active layer over an intermediate time period.

The hydrological regime of the sites was monitored by lysimeters and neutron moderation. Frost heaving at the sites was measured with telescoping heave tubes (Mackay et al. 1979) and by the differential movement of magnets buried within the soil profile. The ground thermal regime was monitored with thermistor cables. The process of ice lens growth at the base of the active layer was examined with PVC tubes 2 m long filled with soil slurry and placed in the ground.

Additionally, two ground plots were irrigated with tritium-rich water in 1983. These plots were core drilled in July 1984 and 1985. The tritium content profiles of the cores were examined in search of independent evidence for water movement into permafrost on an annual basis.

The experiments were initiated on 1 August 1983, and are reported here up to 8 October 1985. Measurements were made at frequent intervals throughout the period.

1.6 Objectives and organization.

The objectives of this research may be summarized as:

- (1) to document the mass flux of water through the surface layers of permafrost terrain on an annual basis.
- (2) to assess the rate of movement of water into permafrost over the past 35 years through analysis of the tritium content of ground ice.
- (3) to examine the stratigraphy and oxygen-isotope ratios of ground ice in the Mayo area for support of the hypothesis on a geological time scale.
- (4) to determine the rate of permafrost aggradation at the sites in terms of the rate of growth of aggradational ice.
- (5) to deduce the hydrological properties of soils in the study area at temperatures not practicable in the laboratory (following Smith 1985a).

In addition, the Quaternary geology of the region has been reviewed in order to provide an appropriate background to the study of ground ice stratigraphy in the area and a geomorphological context within which to place the analysis of present processes.

The thesis is presented in six chapters. The next chapter provides a review of the geotechnical literature pertaining to the movement of water in frozen soil and the mass balance of permafrost. The hypothesis of this thesis is developed in detail. Following this, a description of the study area, the study sites

and experimental techniques is given. In Chapter Four, the mass balance of the sites is computed, and the annual hydrological cycle is discussed in detail. Chapter Five contains an evaluation of the supporting work (tritium profiling and ground ice stratigraphy), and a discussion of the long-term data and short-term results. The work is summarized and concluded in Chapter Six.

CHAPTER 2

HYDROLOGICAL REGIME OF PERMAFROST

Investigations of the surface hydrology of permafrost terrain have conventionally considered the permafrost table as an impermeable barrier (e.g. Brandon 1965; Williams 1970; Rouse 1982a). Streamflow response to snowmelt or rainfall is generally rapid in permafrost regions (Cogley and McCann 1976; Woo and Steer 1982), and subsurface drainage patterns are controlled by variations in active layer thickness (Woo and Steer 1983). Overall, the groundwater contribution to streamflow in the continuous permafrost zone is negligible and leads to an ephemeral regime (Dingman 1973; Slaughter *et al.* 1983); in the discontinuous zone flow is greatly reduced in winter (Brandon 1966; Kane *et al.* 1973).

Conventional hydrologic research assumes that the movement of water in the thawed active layer is dominated by gravity (Harlan 1974). This is a reasonable assumption for the snowmelt period, when the bulk of catchment discharge occurs and the active layer is not deep (Woo *et al.* 1981, 1983). The paradigm is borrowed from southern regions, where seasonal frost supports rapid runoff of snowmelt (Haupt 1967; Waylen 1985), some of which may infiltrate frozen soil (Engelmark 1984; Granger *et al.* 1984). In summer, permafrost watersheds behave similarly to southern examples where impermeable horizons or strata lie close to the surface.

The hydrology peculiar to permafrost terrain is that relating to the movement of water close to the base of the active layer during the thaw season (Mackay 1983a), and throughout the soil profile after freeze-back (Parmuzina 1978; Mackay *et al.* 1979; Mackay and Lewis 1981). This water movement is controlled by pressures generated by the presence of ice and water in soils at temperatures below 0°C , which may be understood by considering the behaviour of fluids in porous media (e.g. Everett 1961; Romkens and Miller 1973).

The first part of this chapter considers the behaviour of water in frozen soils. Subsequently, the annual hydrological cycle of permafrost is described and the development of the ice-rich zone immediately beneath the active layer is discussed.

2.1 Unfrozen water in soils at temperatures below 0°C .

Beskow (1935), Nersesova (1953) and Lovell (1957) were amongst the first to recognise that a considerable proportion of the water in soils at temperatures below 0°C is in the liquid phase. The unfrozen water content (w_u , g g^{-1}) declines with temperature, and is described by the unfrozen moisture content characteristic curve (Figure 2.1), which is unique to a soil (Williams 1967). The presence of unfrozen water is due to several factors, principally the electrochemical effects in pore water associated with the electrical fields emanating from soil particle surfaces (Mitchell 1976); the presence of solutes in soil water (Prigogine and Defay 1954; Banin and Anderson 1974; Hallet 1978);

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

and the restricted volume of soil pores, which prevents ice nucleation at 0°C (Everett 1961). In particular, electrochemical effects are responsible for the persistence of films of unfrozen water around the surfaces of soil particles at temperatures well below 0°C (Low et al. 1968).

The surfaces of soil particles are sites of unbalanced charge, due to isomorphous substitutions in the composition of molecular lattices and the locations of fractures in lattice sheets (Mitchell 1976). The specific surface area (SSA, surface area of particles per unit volume of soil) is the most important factor determining the amount of such charge per unit volume of soil. Of all soil materials, clays possess the highest SSA (Hillel 1971), and the clay content (or SSA) of a soil may be used as an indication of w_u (Dillon and Andersland 1966; Anderson and Tice 1972).

Water molecules are attracted to the negatively charged surfaces of soil particles, where they are prevented from orientating into an ice lattice by hydrogen bonding (Figure 2.2a). In addition, cations are attracted to the charged surfaces, thereby increasing the osmotic potential in the pore water and lowering the freezing point (Figure 2.2c). Hydration occurs between molecules and the cations in solution, inhibiting ice formation (Figure 2.2b). Finally, dipolar attraction accounts for some unfrozen water held close to the clay surface (Figure 2.2d) (Mitchell 1976).

The nature of the surface-charge interaction has been

Figure 2.2. Mechanisms of water adsorption by soil particle surfaces. (a) Hydrogen bonding. (b) Ion hydration. (c) Attraction by osmosis. (d) Dipole attraction. (Mitchell 1976, Fig. 6.4).

modelled most successfully by the Guoy-Chapman double layer theory (Mitchell 1976). This considers the charge emanating from a soil particle to decay exponentially from the surface of the particle, at a rate determined by the concentration of dissolved cations in the pore fluid. The double layer of water and ions remains unfrozen below 0°C , but as temperature declines, its thickness is reduced. The double layer accounts for the presence of unfrozen water in soils at temperatures as low as -70°C (Anderson and Morgenstern 1973).

2.2 Pore pressure gradients due to freezing.

Water in frozen soil is subject to several forces that lower its potential (pore pressure), creating a suction relative to bulk water (Hillel 1971; Williams 1982). Soil water potential, ψ , is commonly divided into five principal components: gravitational (elevation); matric; osmotic; pneumatic (overburden); and, in freezing soils, electrical potential (Hillel 1971; Harlan 1974; Parameswaran 1982). Both osmotic and electrical potentials may be important locally, particularly during the freezing of saline soils (Hallet 1978; Mahar et al. 1982; Parameswaran and Mackay 1983). However, the matric potential, created by the capillary and adsorptive forces of the soil matrix, is the principal determinant of the pore pressure gradient within the soil (Williams 1982).

It is useful to note the similarity between the pressure state of water in unsaturated soil and in frozen ground (Williams

1964; Koopmans and Miller 1966). As soil dries, water is held in tension on the surfaces of soil particles, and the pore water pressure declines (Hillel 1971). (The relationship between water content and pore water pressure for unsaturated soils is known as the suction-moisture content characteristic.) Whilst air pressure in the pores is close to atmospheric, water pressure is considerably lower, and is maintained by the surface tension of the air-water interface. A suction (matric potential) gradient develops, drawing water towards drier parts of the soil. Water movement along the suction gradient continues in unsaturated soil, although the hydraulic conductivity is greatly reduced (Williams 1982).

In a similar fashion, the unfrozen water in frozen soil is held at high suctions (Koopmans and Miller 1966; Williams 1967). The pressure difference between the ice and water in equilibrium is maintained by the surface tension of the ice-water interface, and is approximately proportional to the temperature depression below 0°C. The relationship between ice and water pressures and temperature is described by the Clausius-Clapeyron equation (Edelfsen and Anderson 1943; Loch 1980):

$$V_w \frac{dP_w}{dT} - V_i \frac{dP_i}{dT} = \frac{L}{T} \quad (2.1)$$

where,

V_i, V_w = specific volumes of ice and water (1.1 and $1.0 \times 10^{-3} \text{ m}^3 \text{ kg}^{-1}$),

P_i, P_w = pressures on the ice and water (Pa),

T = temperature (K),

and L = latent heat of fusion of water ($3.34 \times 10^5 \text{ J kg}^{-1}$).

If we assume that P_w remains constant (ice takes the place of air in the frozen soil/dry soil analogy) then we may rewrite equation (2.1):

$$V_w \frac{dP_w}{dT_w} = \frac{L}{T} \quad (2.2),$$

which, in finite form is:

$$\Delta P_w = \frac{L \cdot \Delta T}{V_w T} \quad (2.3)$$

(note that P_w declines below 0°C , since $\Delta T < 0^\circ$). Hence, near to 0°C (273.15 K), we have:

$$\begin{aligned} \Delta P_w &= \frac{3.34 \times 10^5 \text{ J kg}^{-1}}{1 \times 10^{-3} \text{ m}^3 \text{ kg}^{-1}} \times \frac{\Delta T \text{ K}}{273.15 \text{ K}} \\ &= 12.08 \times 10^5 \text{ J m}^{-3} \text{ K}^{-1} \times \Delta T \text{ K} \\ &= 12.08 \times 10^5 \text{ N m}^{-2} \text{ K}^{-1} \times \Delta T \text{ K} \end{aligned} \quad (2.4)$$

for the change in pressure of w_u with T . This is frequently given as $12.12 \text{ atm } ^\circ\text{C}^{-1}$ or $1.2 \text{ MPa } ^\circ\text{C}^{-1}$ (Edelfsen and Anderson 1943).

An implication of the relationship is that as the temperature of an unsaturated soil drops below 0°C , water does not freeze until the appropriate temperature depression is reached for soil water in tension. The unfrozen water content characteristic curve not only describes the relationship between unfrozen water content of saturated soils and temperature, but also indicates the temperature at which pore water freezing begins in unsaturated soil (Williams 1967).

This temperature-pressure relationship can translate into a strong suction gradient, under natural conditions. For example,

with a temperature gradient of 1°C m^{-1} , a soil water potential gradient of 1.2 MPa m^{-1} ($= 121.1 \text{ m(H}_2\text{O) m}^{-1}$) may be established. However, it is important to note that this is the maximum potential gradient that may develop. In many instances, P_i is not zero (relative to atmospheric pressure), since a growing ice lens may have to overcome considerable resistance offered by the overlying frozen soil (Tsytovich 1973). In these cases, while the difference $P_i - P_w$ is maintained at any given temperature, dP_w/dz declines (Williams and Wood 1985, Wood 1985). Nevertheless, it is apparent that the hydrological regime of permafrost terrain is implicitly linked to the thermal regime.

2.3 Hydraulic conductivity of frozen soil.

The first experimentally determined values for the hydraulic conductivity, K_f , of frozen soil were provided by Williams and Burt (1974; Burt 1974; Burt and Williams 1976), who established that K_f varies with soil type and declines with temperature. The curves of K_f against T obtained by Burt and Williams (1976) vary with soil type similarly to those of w_u (Figure 2.3). K_f was explained in terms of w_u (Burt and Williams 1976, p. 349), but it was also reported that the development of ice lenses across a sample perpendicular to the pressure gradient did not prevent water movement. Similar results were obtained subsequently by Mageau and Morgenstern (1980).

The values of K_f reported by Burt and Williams (1976) were obtained by applying a pressure gradient across a permeameter in a

Figure 2.3 Hydraulic conductivities of a range of frozen soils. (Burt and Williams 1976, Fig. 6).

lactose bath, observing flow through the permeameter at various temperatures and determining K_f from the Darcy equation. The values varied from 10^{-6} to $10^{-11} \text{ m s}^{-1}$, for the temperature range -0.01° to -0.4° C . However, Horiguchi and Miller (1980, 1983) and Loch and Kay (1978) determined values for K_f which were several orders of magnitude lower for the same temperatures, from water flow induced by a temperature gradient. These values are, perhaps, more representative, since both Perfect (1980) and Williams and Wood (1981) also failed to replicate Burt and Williams' (1976) values, due to difficulties with experimental procedure. The values for K_f of 10^{-8} to $10^{-12} \text{ m s}^{-1}$ presented by Loch and Kay (1978), Horiguchi and Miller (1980, 1983) and Williams and Wood (1981) for a range of fine-grained soils at temperatures between -0.01° C and -0.3° C (Figure 2.4), compare with values of 10^{-6} to $10^{-12} \text{ m s}^{-1}$ for the permeability of unfrozen glacial till or marine clay (Freeze and Cherry 1979).

Smith (1985a) has presented determinations of K_f for a silty clay soil at temperatures as low as -2.4° C . The values were obtained by comparing field observations of frost heave with potential gradients derived from ground temperatures recorded simultaneously. The determinations ($2 \times 10^{-11} \text{ m s}^{-1}$ at -0.5° C , $3.3 \times 10^{-12} \text{ m s}^{-1}$ at -1.0° C , $10^{-12} \text{ m s}^{-1}$ at -2.4° C , Smith 1985a, Table 3) are conservative estimates of K_f for two principal reasons:

(i) Frost heave is the result of a flux divergence in a layer of soil, with water flowing into the layer and freezing, and may

Figure 2.4. Hydraulic conductivities of several fine-grained materials. (Horiguchi and Miller 1983, Fig. 2).

therefore be an underestimate of the water flow through a layer of soil.

(11) $d\psi/dz$, may be less than $1.2 \text{ MPa} \times dT/dz$ (see above), and, hence, K_f will be underestimated.

These results, and others presented by Oliphant et al. (1983), suggest that $10^{-13} \text{ m s}^{-1}$ may be a reasonable lower boundary for the range of K_f with temperature (Smith 1986).

2.4 Regelation.

The apparent permeability of ice lenses and the ability of soil particles to move through ice, has been discussed by Miller (1970) and Römken and Miller (1973). Miller et al. (1975) adapted the concept of regelation, commonly encountered in glaciology (Sugden and John 1976; Patterson 1981), for application to frozen soils. Regelation involves the transfer of water through ice masses in frozen soil by pressure-induced freezing on the upstream (warm) side of the ice and thawing on the downstream (cold) side. Miller et al. (1975, p. 1031) summarized the phenomenon:

"The Clapeyron equation (see equation (2.1)) indicates that when $dT < 0^\circ\text{C}$, P_i exceeds P_w . When, on one side, P_w is increased somewhat, the equilibrium temperature rises above ambient, favouring the formation of ice at that side, and the ice pressure rises. As the increase in ice pressure is felt at the other side, where P_w has not changed, equilibrium temperature falls below ambient, favouring melting. The result is a translational movement of ice... ."

The model of water movement in freezing soils that emerges from these discussions is of a series-parallel process, with series (regelation) flow through ice lenses and crystals in soil pores, and parallel transport in unfrozen water films around the sides of soil particles. Horiguchi and Miller (1980) reported that the effective K_f of ice lenses between -0.05°C and -0.15°C was higher than that of frozen soil ($10^{-10} \text{ m s}^{-1}$ compared to $10^{-11} \text{ m s}^{-1}$), but varied considerably less with temperature. Similar results were obtained by Wood and Williams (1985a) at temperatures as low as -0.3°C (Figure 2.5).

2.5 Vapour diffusion.

Vapour diffusion in soils is driven by the gradient in soil water vapour pressure, de/dz (Giddings and LaChapelle 1962). In frozen soils, the appropriate vapour pressure is the saturation vapour pressure over ice, since w_u is in thermodynamic equilibrium with pore ice. Furthermore, vapour flow in a snowpack overlying the soil may be regarded as a continuation of vapour flow in the soil, since the snow is a vapour-saturated porous medium, with vapour pressure controlled by the ice phase (Powers, et al. 1985).

The rate of vapour flow, Q_v , can be determined by (Giddings and LaChapelle 1962):

$$Q_v = \frac{-D(w)_v}{R T} \cdot \frac{de}{dz} \quad (2.5)$$

Figure 2.5 Relationship of the transport coefficient of ice to temperature. (Wood and Williams 1985a, Fig. 3).

where e = saturation vapour pressure over ice (J m^{-3}),
 R = gas constant for water vapour ($0.46 \text{ J g}^{-1} \text{ K}^{-1}$),
 $D(w)_v$ = diffusivity of the vapour phase ($\text{m}^2 \text{ s}^{-1}$),
 and the other terms are as defined earlier. Since e is a function of temperature, we may write:

$$Q_v = -\frac{D(w)_v}{R T} \cdot \frac{de}{dT} \cdot \frac{dT}{dz} \quad (2.6).$$

Santeford (1978) and Woo (1982) measured apparent vapour movement out of the soil over the winter near Fairbanks, Alaska, and Resolute, N.W.T., respectively. In each case, the vapour flow out of the ground was seen as a continuation of movement within the soil, due to the saturation vapour pressure gradient driving water vapour towards lower temperatures. Total outward fluxes of over 30 mm and 3 mm were measured at the sites over the period, with the lower amount in the frigid conditions at Resolute. However, a close examination of the data suggests that such flows may not be due to vapour transport alone, but ~~liquid~~ migration through the soil coupled with evaporation at the soil surface (Smith and Burn, in press).

2.6 Relative importance of movement in the different phases.

Since the phase composition of soil water varies with temperature, we may also expect the relative importance of the transport mechanisms, i.e. vapour diffusion, liquid flow, and regelation transport, to vary with temperature. Both liquid and vapour flow will occur in ice-free frozen soils (Nakano et al. 1983a). In unsaturated frozen soils with ice present, we may

anticipate liquid flow as controlled by the Clapeyron equation, vapour flow governed by the pressure gradient of saturated vapour over ice, and some regelation transport in pore ice (Miller *et al.* 1975), or across small ice lenses (Oliphant *et al.* 1983). In apparently saturated soil, vapour diffusion may occur in small air pockets trapped in the soil.

2.6.1 Water movement in saturated soils.

A critical observation made by Burt and Williams (1976, p. 355) regarding water movement in saturated soils is that "the presence of ice lenses in a sample reduces its hydraulic conductivity compared to an unlensed sample at all temperatures." Data supporting this statement, however, were only presented for the temperature range 0°C to -0.1°C , where the presence of ice lenses in samples of Carleton silt reduced K_f by two orders-of-magnitude (Burt and Williams 1976, Fig. 7). Both Loch and Kay (1978) and Horikuchi and Miller (1983) obtained values for K_f of between 10^{-8} and $10^{-11} \text{ m s}^{-1}$, for a range of fine-grained soils between 0°C and -0.1°C . In the warmer half of this range, the transport coefficient (K_f) of ice ($10^{-10} \text{ m s}^{-1}$, Wood and Williams 1985a) is less than K_f of bulk soil (10^{-8} to $10^{-10} \text{ m s}^{-1}$), and therefore ice impedes the transport of water through the soil.

However, ice lenses are formed at temperatures below 0°C (Loch and Kay 1978; Miller 1978; Konrad and Morgenstern 1980, 1982), and during soil freezing the bulk of water transport between the 0°C isotherm and the first ice lens is in the liquid

phase. This zone may be divided into two parts: a warm part, between 0°C and the warmest pore ice, where water is still liquid; and a cold part, between the warmest pore ice and the ice lens, known as the "frozen fringe" (Figure 2.6) (Miller 1978, O'Neill and Miller 1985). In this region, liquid transport occurs either in the unfrozen pore water or in the adsorbed water surrounding particles in the ice, expelling the particles from the growing lens (Römken and Miller 1973; O'Neill and Miller 1985). Loch and Kay (1978) and Konrad and Morgenstern (1980) have presented data which indicate that the first ice lenses form in silty soils at temperatures between -0.05 and -0.3°C .

At lower temperatures, where both ice and water coexist, measurements of the transport coefficient of the ice phase alone (10^{-11} to $10^{-12} \text{ m s}^{-1}$, Wood and Williams 1985a), are up to two orders-of-magnitude higher than the hydraulic conductivity of bulk soil (Loch and Kay 1978; Horiguchi and Miller 1983). It is reasonable to suppose, therefore, that regelation does not inhibit moisture flow below about -0.1°C . Instead, the slow movement in liquid films away from the cold side of the lens reduces the pressure difference ($P_i - P_w$) generating regelation transport. Thus, the critical phase determining K_f is the liquid phase.

2.6.2 Water movement in unsaturated soil.

In unsaturated soil, water moves by both vapour diffusion and liquid flow in adsorbed water films. Jackson (1965) estimated that vapour diffusion accounts for 60 to 80 percent of total water

Figure 2.6 Schematic diagram of the frozen fringe, with ice lens above. (O'Neill and Miller 1985, Fig. 1).

Arrows indicate direction of mass transport.

movement in very dry soil (0.02 g g^{-1} soil), but since it is difficult to isolate the phases experimentally, water movement in unsaturated soils is generally analysed with equations of the form (Nakano et al. 1984):

$$Q = -D(w) \cdot \rho_w \cdot \frac{\partial w}{\partial z} \quad (2.7)$$

where,

Q = flow ($\text{kg m}^{-2} \text{ s}^{-1}$),

$D(w)$ = soil-water diffusivity ($\text{m}^2 \text{ s}^{-1}$),

ρ_w = dry density (kg m^{-3}),

and w = weight of water per unit weight of dry soil.

$D(w)$ accommodates the flow component in both vapour and liquid phases; it increases with w , and is related to the hydraulic conductivity of a soil, K , by:

$$D(w) = K \cdot \frac{d\psi}{dw} \quad (2.8).$$

The term $d\psi/dw$ represents the slope of the suction-moisture content characteristic curve (Williams 1982), which is analogous to the unfrozen water content characteristic curve (Koopmans and Miller 1966). Indeed, since

$$\frac{d\psi}{dw} = \frac{d\psi}{dT} \cdot \frac{dT}{dw} \quad (2.9),$$

and,

$$\frac{d\psi}{dT} = 1.212 \text{ MPa } ^\circ\text{C}^{-1} \quad (2.10)$$

in frozen soil, $d\psi/dw$ may be obtained from the slope of the unfrozen water content characteristic curve, dT/dw .

Values for $D(w)$ determined by Nakano et al. (1982) with

isothermal, ice-free, frozen soils at low water contents ($10^{-9} \text{ m}^2 \text{ s}^{-1}$) are comparable to those obtained by Jackson (1964, 1965) for similarly dry unfrozen soils. With this in mind, Nakano *et al.* (1983a, p. 892) concluded that "the transport mechanism of water in soil containing no ice at subzero temperature is essentially the same as that in unfrozen soil containing a small amount of water".

However, both Jackson (1965) and Hillel (1971) indicate that at water contents above about 0.04 g g^{-1} , $D(w)$ increases rapidly with w , due to liquid diffusion. Both of these writers imply that vapour diffusion is only important at very low w .

It is interesting to note, therefore, that values for K_f determined from Nakano *et al.*'s (1982) data for $D(w)$, using equation 2.8, are similar to determinations of K_f at the same temperature made by Smith (1985a) and Ohrai and Yamamoto (1985) for saturated systems. (The hydraulic conductivity of unsaturated frozen soil may be estimated from equation 2.8 using determined values of $D(w)$ and $d\psi/dw$. For example, at -1.0°C , $D(w) = 10^{-9} \text{ m}^2 \text{ s}^{-1}$ for Morin clay (Nakano *et al.* 1982), $dT/dw = 33^\circ\text{C g g}^{-1}$ (Nakano *et al.* 1982, Fig. 1), $d\psi/dT = 1.2 \text{ MPa } ^\circ\text{C}^{-1}$ and, therefore, $K_f = 1.5 \times 10^{-13} \text{ m } \cdot \text{s}^{-1}$ (Smith and Burn, in press).) Indeed, Nakano *et al.* (1982) determined maximum values for $D(w)$ at a water content equal to the unfrozen water content of the soil. Together, these results suggest that most water movement in frozen soil is in the liquid phase.

A quantitative estimate of the relative importance of vapour

flow in frozen soil can be obtained when results from equation 2.9 are compared to estimates of flow rates in the liquid phase using Darcy's Law (Oliphant et al. 1983; Smith 1985a). For given temperature conditions in the soil, the liquid flux, Q_l , is given by (ignoring signs):

$$Q_l = K_f \cdot \frac{d\psi}{dT} \cdot \frac{dT}{dz} \quad (2.11).$$

At -1°C , for example, K_f may be approximately $10^{-12} \text{ m s}^{-1}$ (Smith 1985a), and $d\psi/dT$ is $1.212 \text{ MPa } ^\circ\text{C}^{-1}$. Thus,

$$\begin{aligned} Q_l &= 10^{-12} \text{ m s}^{-1} \times 1.2 \text{ MPa } ^\circ\text{C}^{-1} \cdot \frac{dT}{dz} ^\circ\text{C m}^{-1} \\ &= 1.2 \times 10^{-10} \cdot \frac{dT}{dz} \text{ m s}^{-1} \end{aligned}$$

Under the same conditions, vapour flow rate, Q_v may be estimated by:

$$Q_v = \frac{D(w)}{R T} \cdot \frac{de}{dT} \cdot \frac{dT}{dz} \quad (2.10, \text{ above}).$$

If $D(w)$ with respect to vapour is $0.4 \times 10^{-9} \text{ m}^2 \text{ s}^{-1}$ (Jackson 1965; Nakano et al. 1982), and de/dT is the slope of the saturation vapour pressure curve at -1.0°C ($48 \text{ Pa } ^\circ\text{C}^{-1}$, Weast and Astle 1982), we have, under optimum conditions:

$$\begin{aligned} Q_v &= \frac{0.4 \times 10^{-9} \text{ m}^2 \text{ s}^{-1}}{1.2 \times 10^2 \text{ J } ^\circ\text{C}^{-1}} \times 48 \text{ Pa} \times \frac{dT}{dz} \\ &= 1.4 \times 10^{-16} \frac{dT}{dz} \text{ m s}^{-1}. \end{aligned}$$

Clearly, vapour flow is several orders-of-magnitude less than liquid movement, even with high values for $D(w)$. It may be neglected for most practical purposes.

2.7 Coupled heat and moisture flow in ground below 0°C.

Darcy's Law represents the flow of water in porous media as the product of hydraulic conductivity and a potential gradient. Burt and Williams (1976) measured flow through frozen soils under an applied pressure gradient, and computed K_f as the ratio of pressure gradient to observed flow. More recently Perfect and Williams (1980), Yoneyama et al. (1983), Ohrai and Yamamoto (1985) and Wood and Williams (1985b) have measured the movement of water in frozen soil in response to a temperature gradient. In these cases, the temperature gradient has been converted to a potential gradient using equation 2.4.

The rate of movement of water in frozen soil, however, is not only dependent on the hydrological characteristics of the soil, but also on the thermal properties. Water movement to regions of lower temperature results in freezing and the release of latent heat, which must then be dissipated for the temperature gradient to remain constant. (Recall the necessity for water to freeze as prescribed by the unfrozen water content characteristic curve.) Hence, the flow of water in frozen soil is governed by a coupled equation:

$$C_a \frac{\partial T}{\partial t} - \frac{\partial}{\partial z} \left(k \frac{\partial T}{\partial z} \right) = L \rho_w \frac{\partial}{\partial z} \left(K_f \frac{\partial \psi}{\partial z} \right) \quad (2.12)$$

where, $C_a = C_m + L_w \cdot \partial \theta_u / \partial T$

and, k = thermal conductivity ($W m^{-1} K^{-1}$),

C_m = volumetric heat capacity of the soil-water-ice mixture ($J m^{-3} K^{-1}$),

θ_u = volumetric unfrozen water content ($cm^3 cm^{-3}$),

t = time (s),

and the other terms are already defined (Smith 1985b, after Harlan 1973). The first term on the left hand side of the equation indicates heat released from the soil constituents and by water frozen in situ as the system cools; the second term describes heat conducted through the soil. The right hand side of equation 2.12 represents the latent heat released by the water which moves into frozen soil and freezes there (see Figure 2.6).

The equation is valid if the convective heat conveyed by the liquid water to the frozen soil is negligible. Since the heat capacity of water per Kelvin is almost two orders-of-magnitude lower than the latent heat of fusion, and the water is rarely more than a few tenths of a degree warmer than the frozen fringe, the convective component can safely be ignored.

Except for K_f and k , the terms on each side of equation 2.12 behave similarly as temperature varies. However, whereas K_f declines sharply with temperature (Burt and Williams 1976), k is relatively constant (Kay et al. 1981). Therefore, if k is a limiting property of the system--e.g. when gentle temperature gradients are present--water migrates slowly to a given position in the soil and freezes, forming an ice lens. In this case, the hydraulic conductivity of the system does not restrict water supply. On the other hand, with a steep temperature gradient, K_f acts as the limiting property, the supply of moisture to the frost front is impeded, and the 0°C isotherm can penetrate the soil.

An example of the interrelationship between heat and moisture flow has been provided by Harlan (1974), who considered the hydrogeological conditions at the base of permafrost. Under conditions of hydrostatic equilibrium, any difference in elevation between two points must be off-set by an equal but opposite change in water potential. Using a form of equation 2.3, Harlan (1974), calculated that hydrostatic equilibrium would occur with a vertical temperature gradient of $0.00816^{\circ}\text{C m}^{-1}$, a value too low to be commonplace. (A geothermal gradient of $0.023^{\circ}\text{C m}^{-1}$ has been measured in southern and central Yukon (Burgess *et al.* 1982), a factor of three higher than the hydrostatic equilibrium gradient.) Given an adequate water supply, we may expect, therefore, a constant upward migration of groundwater into the base of permafrost. However, this may not occur at rates predicted simply by calculating $d\psi/dz$ from the temperature gradient, since the overburden pressure at the base of permafrost may be considerable (perhaps of the order of 0.1 to 1 MPa), and P_i will be elevated above atmospheric pressure. In this case, since the difference $P_i - P_w$ remains constant for any temperature, $d\psi/dz$ is reduced across the unfrozen/frozen soil boundary.

2.8 Frost heave.

The growth of ice lenses usually results in heaving of the soil. Frost heave is defined as the volumetric expansion that occurs when water is drawn to regions of the soil which become supersaturated with ice (Williams 1982). Frost heave is dist-

inguished from the nine percent volume expansion that occurs as water freezes, and is restricted to soil expansion due to water migration. The factors affecting frost heave have been studied for some time, e.g. Taber (1929, 1930), Beskow (1935), but only recently have the mechanics and thermodynamics of frost heave been considered together (e.g. Miller 1972, 1978; Harlan 1973; Gilpin 1980; O'Neill and Miller 1985; Williams and Wood 1985b).

For the purposes of this thesis, two frost heave regimes are defined. Primary heave is the soil displacement at the warmest (primary) ice lens, resulting from water movement in the liquid phase or by regelation through the frozen fringe. The frozen fringe thickens as the frost front penetrates the soil and the ice lens grows at ever colder temperatures, until, at some point, the thermal, hydrological and mechanical properties of the system favour the formation of a new lens closer to 0°C . At that time, the location of ice lens growth jumps to the new site. This mechanism accounts for the repeated banding of ice lenses often observed in frozen soils.

The thickness of each primary ice lens depends on the speed of penetration of the frost front. As the rate of propagation of the frost front declines, usually with depth, more time is available for the growth of each lens before a new one is initiated. The limiting case of primary heave occurs when the rate of propagation of the freezing front is zero, and the rate of heat extraction from the system equals the rate of freezing of water arriving at the primary ice lens. This frost heave regime has

been described in detail by Miller (1972, 1978), Loch and Miller (1975), Loch (1980) and O'Neill and Miller (1985).

Secondary heave is used here to refer to heave due to long-term water redistribution within frozen ground behind the primary ice lens. In this case, the primary lens acts as a reservoir for water molecules entering the soil-water system in frozen ground by regelation. Goto and Takahashi (1982), Yoneyama *et al.* (1983), Ohrai and Yamamoto (1985) and Smith (1985a), amongst others, have shown that soil displacement may occur due to the redistribution of moisture on the cold side of the primary ice lens, even with a stationary frost front. This takes place at a slow rate, partly because K_f is low, about $10^{-12} \text{ m s}^{-1}$ (Burn and Smith 1985a; Smith 1985a). In addition, the resistance of frozen soil to displacement by a growing ice lens raises the ice pressure, and, therefore, the soil water pressure gradient may be reduced (Wood 1985).

The processes generating primary heave are generally understood, and are incorporated in various models and engineering tests (e.g. Gilpin 1980; Konrad and Morgenstern 1980, 1981, 1982, 1983). However, secondary heave remains a phenomenon of disputed significance (Williams 1979). Mageau and Morgenstern (1980, p. 59) expressed the opinion that "regelation (and, hence, secondary heave)...is very slow and would probably contribute little to the total amount of frost heave over an engineering time frame." Their conclusion, however, was based on tests which lasted for only a few days; some of the results in this thesis suggest

otherwise.

2.9 Annual water balance of permafrost.

It is convenient to divide the hydrological year of permafrost terrain into four periods: spring snowmelt, summer thaw, fall freeze-back and winter.

2.9.1 Spring snowmelt.

Snowmelt occurs as the snowpack is warmed by advected air masses or local heating of exposed ground and tree surfaces. Percolation and refreezing of meltwater lower in the snowpack contributes to the warming of the snowpack and the development of basal ice layers (Woo and Heron 1981; Marsh and Woo 1984).

The latent heat released by the refreezing of meltwater brings the snowpack to an isothermal state at 0°C fairly quickly (Marsh and Woo 1984), and establishes conditions favouring temperature-induced infiltration of meltwater into the underlying, cold soil (Figure 2.7) (Kane and Stein 1983a; Burn and Smith 1985a). Snowmelt may infiltrate the ground by gravity flow down desiccation and thermal contraction cracks (Mackay 1974a, 1984; Kane 1980; Kane and Stein 1983a, 1983b), but this process is limited by water freezing rapidly in the frozen soil and blocking the cracks. Such infiltration and refreezing contributes to some warming of the soil (Woo and Heron 1981; Wright 1981; Kane and Stein 1983a).

Kane (1980) and Kane and Stein (1983a) have also shown that

Figure 2.7. Snow-soil temperatures during snowmelt in 1981,
at a site near Resolute, N.W.T. (Marsh and Woo
1984, Fig. 7).

infiltration may continue at a constant rate for periods of days, and that most groundwater recharge in parts of Alaska with seasonal frost occurs during snowmelt, before the ground is thawed (Kane and Stein 1983b). Interestingly, their data show an almost constant rate of infiltration about thirty minutes after tests began; this can be explained in terms of temperature-induced water flow into the frozen soil (Burn and Smith 1985b).

In the snowmelt period, "warm" soil temperatures and abundant moisture create thermal and hydrological conditions optimal for downward water movement into frozen ground. By the end of snowmelt, the surface layers of the ground are near 0°C (Woo and Heron 1981; Kane and Stein 1983a), and thawing of the active layer begins almost as soon as the ground is exposed.

2.9.2 Summer thaw.

The thawing of the active layer is accompanied by migration of water from thawed soil into the frozen ground below (Figure 2.8). The flow is induced by the ground temperature gradient, and is supplied by melt from lingering snow patches, precipitation and ice lenses thawing in the soil (Wright 1981; Lewkowitz and French 1982; Mackay 1983a).

Wright (1981) observed sudden warming of frozen ground just below the thawing front, which he attributed to infiltration and freezing of water from the thawed layer. Ground temperatures below the thawing front rose to nearly 0°C some time before the relevant soil layers were thawed, indicating that there was

Figure 2.8 Vertical water movement in the active layer and surface of permafrost. (Mackay 1983a, Fig. 8).

"significant penetration of suprapermafrost groundwater into the still-frozen parts of the active layer" (Wright 1983; p. 1416). Further, Mackay (1983a, Table 1) has presented measurements of heave made on Garry Island, N.W.T., and at Illisarvik and Inuvik, N.W.T., which indicate soil displacement below the thawed active layer due to downward water migration. Thaw season heave of up to 3.7 mm, 2.9 mm and 9.0 mm, respectively, was measured at the sites.

When the active layer reaches its maximum depth, water is drawn into permafrost by the same process. Cheng (1983) reported water content increases of up to $0.05 \text{ cm}^3 \cdot \text{cm}^{-3}$ in the surface layers of permafrost at Fenghuoshan, P.R.C., in the period 22 May to 2 October, 1976 (Figure 2.9). Parmuzina (1978) reported an increase of 10 to 25 percent in the gravimetric water content of the still frozen active layer during summer, and observed a similar increment in the upper permafrost zone. Further, Mackay and Lewis (1981) have reported observations, made at Inuvik over several years, of slight heave of three bench marks buried to two metres. This long-term heave may be due to a similar process.

Parmuzina (1978) and Cheng (1983) noted that the downward flux of water into frozen ground leads to the development of "horizontal streaks" of ice lenses in the frozen soil. Parmuzina (1978, p. 145) also observed that "the cryogenic texture of the upper part of permafrost was a continuation of the texture of the streaky horizon in the active layer". This cryotexture was referred to by Cheng (1983, p. 57) as "thick-layered ground ice",

1. May 2
2. Oct. 2
3. Maximum depth of seasonal thawing

Figure 2.9 Soil water content change over summer 1976,
at Fenghuoshan, P.R.C. (Cheng 1983; Fig. 7).

and Parmuzina (1978) noticed that the layers of ice thickened with depth below the permafrost table. Both of these writers acknowledged that the soil just above the thawing front was at, or near, saturation due to the water released by the thawing of ice lenses, and that conditions were suitable for downward water migration.

2.9.3 Fall freeze-back.

Once ground surface freezing starts, ground temperatures fall rapidly (e.g. Wright 1981; Rouse 1984; Smith 1985a). The active layer freezes from both top and bottom, but the rate of upward freezing is small (Parmuzina 1978; Cheng 1983; Smith 1985a). (The total component of upwards freezing may be overestimated by frost tubes that utilise the freezing of bulk water to monitor the movement of the frost front (e.g. Mackay 1973b), since the frost front is not located at 0°C , particularly in fine-grained soils.) The slower upward freezing is to be expected since the surface temperature gradient is usually steeper than the gradient at depth (e.g. Smith 1985a, Fig. 9), and hence heat flow upward is more rapid. The penetration of the frost front into the active layer leads to moisture redistribution within the soil, principally upward to form segregated lenses in the upper decimeters (Parmuzina 1978; Mackay et al. 1979; Mackay 1980; Smith 1985a). By the end of freeze-back the lower part of the active layer is relatively desiccated (Parmuzina 1978; Smith 1985a).

Net upward movement of water is dependent upon the rate of propagation of the frost front into the soil (Cheng 1983), but

continues within frozen soil (Mackay *et al.* 1979; Smith 1985a). Smith (1985a) measured heaving of frozen parts of the active layer during freeze back, calculated values of K_f for the soil at various temperatures below 0°C , and obtained results generally consistent with laboratory determinations of K_f for similar soils. Both Parmuzina (1978) and Smith (1985a) suggested that moisture availability is the critical factor limiting frost heave in the active layer. The observed heave rates declined as the frost front penetrated the active layer and unfrozen soil became desiccated (Parmuzina 1978; Mackay 1983a; Smith 1985a).

There may also be some water movement downward from the lower regions of the active layer into permafrost or refrozen soil as the frost table rises (Parmuzina 1978; Cheng 1983; Mackay 1983a; Smith 1985a). Moisture continues to flow downward from the base of the unfrozen soil until a monotonic temperature gradient is established, sometime after the active layer has frozen. Wright (1981) observed that temperatures between -0.03°C and -0.08°C persisted for up to three weeks during the freezing of the active layer at sites near Schefferville, P.Q., and reported downward movement of moisture during the same period.

2.9.4 Winter.

For the purposes of this discussion, winter is defined as the period from the completion of freeze-back until the beginning of snowmelt, when the ground temperature profile is monotonic, decreasing upwards.

Although, conventionally, the hydrological regime of the active layer has been regarded as static during the winter (Dingman 1973, Slaughter et al. 1983), results cited (from Chistotinov 1973) and presented by Parmuzina (1978) indicate that this is not the case. Parmuzina (1978) reported an increase in water content in the surface 30 cm of the soil over the period December to late April, accompanied by a slight drop in the moisture content of lower horizons. She attributed this to "either...lateral migration during irregular freezing of the active layer", since she was using a destructive core sampling method, "or...inflow of unfrozen water from the underlying permafrost" (Parmuzina 1978, p. 138). Subsequent analysis of heave measurements led her to reject the first alternative. However, there is no discussion of the contribution of snowmelt to the hydrological regime of the sites, and the small increments at the surface may be partly accounted for by infiltration.

Observations of frost heave at Inuvik reported by Mackay and Lewis (1981) indicated that up to 50 percent of total heave in the surface 2 m of the ground occurred between December and May. In addition, they reported that a total heave greater than the in situ expansion of the water content of the active layer was regularly observed over the five winters between 1975 and 1980. Together, these results suggest "that there is an upward movement of pore water through permafrost during the winter months" (Mackay and Lewis 1981, p. 117).

Similar observations reported by Mackay et al. (1979) show

ground surface heave of 1 to 2 cm between January and May 1978. The amount of heave increased with active layer depth, and was attributed to migration of "substantial amounts of pore water ... from adjacent areas" (Mackay et al. 1979, p. 405). The soil at the Inuvik research site is a fine-grained colluvium (50 percent < 0.002 mm, Mackay et al. 1979), with a considerable proportion of soil moisture unfrozen at -2°C (Smith 1985a, Fig. 4). In contrast, direct observations of moisture content profiles made with a neutron probe at a sandy site near Churchill, Manitoba, by Rouse (1984), did not reveal any change in moisture distribution over the period January to March 1980.

Santeford (1978), Woo (1982) and Marsh and Woo (1984) have reported measurements of vapour flow out of frozen soil over the winter. Santeford (1978), working near Fairbanks, Alaska, covered several sites with a vapour barrier, and measured the water equivalent of the snowpack over the barrier and adjacent (uncovered) ground. He attributed the difference of 30 mm to vapour outflow from the peat soil, and, indeed, observed a similar desiccation in the surface layers of the peat beneath the "natural" snowpack.

Woo (1982) and Marsh and Woo (1984) placed vapour barriers 3 cm above the soil surface at sites near Resolute, N.W.T., and determined the water equivalent of hoar frost formed over the winter on the soil side of the barrier (3 mm). The moisture was assumed to have moved through and out of the soil as vapour.

Similar results (up to 4 mm water equivalent) have been

obtained from sites near Mayo (Smith and Burn, in press). At these sites, ground temperature profiles were monitored closely in order to determine $D(w)_v$, using equation 2.6. The values obtained for $D(w)_v$ were four orders-of-magnitude higher than those obtained by Nakano et al. (1982, 1983b) for soils at similar temperatures. However, when the outflow measurements were combined with temperature data using equation 2.13, a value of $10^{-13} \text{ m s}^{-1}$ was obtained for K_f . This value is consistent with others determined at comparable temperatures (e.g. Oliphant et al. 1983; Smith 1985a).

Since Nakano et al.'s (1982, 1983b) values are similar to other determinations of $D(w)_v$ (e.g. Jackson 1964, 1965), Smith and Burn (in press) have suggested that the moisture moves as liquid through the frozen soil and evaporates at the soil surface. Desiccation of soil layers was not detected in neutron probe measurements made over the winter, as the resolution of the instrument was inadequate to detect the small total loss of moisture from the soil column. However, such desiccation would have been noticed in any one 15 cm layer, suggesting that the soil water supplied to the snowpack was replenished from below.

The net result of water movement in the winter period is to reinforce the variations in the moisture content profile established at the end of freeze back. Parmuzina (1978, Fig. 1) indicated that the water content of the surface 40 to 50 cm of loamy soils increased during both freeze back and winter, while underlying material dried. Desiccation of the uppermost 10 cm of

permafrost was measured from December on. Results presented by Smith and Burn (in press) indicate that upward water movement continues in the soil throughout the winter. The flows upward along the temperature gradient in winter are sufficient to result in water accumulation and frost heaving in the surface layers of the soil (Parmuzina 1978; Mackay et al. 1979; Wright 1981; Cheng 1983).

2.10 Aggradational ice.

The association of the hydrological regime of permafrost with the thermal regime has been made explicit. The thermal gradient governs the hydraulic gradient for much of the year in the surface layers of the ground, although the flux of water may be small due to the magnitude of K_f . (On an annual basis, total water movement may be dominated by the component in the thawed active layer (Wood and Steer 1983).)

While K_f may change by three or four orders of magnitude between 0°C and -0.4°C (Burt and Williams 1976; Horiguchi and Miller 1983), the magnitude of the potential gradient does not change as a function of temperature (Cheng 1983). (Extremely steep gradients, of the order of hundreds of $^{\circ}\text{C m}^{-1}$, or more, would be required to alter the potential gradients in the same proportion.) Hence a flux divergence is to be expected between the (relatively large) water movement downward into permafrost in summer, when ground temperatures are warm, and the (relatively small) upward flow out of this layer under the frigid winter

regime. The result of this imbalance in seasonal fluxes is a net annual accumulation of water in a layer at the top of permafrost "that is usually ice-rich and not desiccated" (Mackay 1983a, p. 121).

An ice-rich zone is indeed frequently observed just below the base of the active layer in permafrost regions (Mackay 1971b, 1972, 1983a; Brown et al. 1978; Outcalt 1982). Conventionally, the origin of the ice has been ascribed to water that has been trapped at the base of the active layer as the permafrost table has risen, due to climatic cooling, the addition of surface material by sedimentation or soil creep, or to vegetational (peat) accumulation (Mackay 1970, 1971b, 1972, 1973b; Brown et al. 1978). Vegetational changes per se may also alter the energy balance at the ground surface, leading to a lower surface temperature and higher permafrost table.

The excess ice content in this zone has been thought to originate from "subsurface horizontal tension cracks....becoming infilled by surface water moving downwards through vertical cracks" (Mackay 1971b, p. 26). However, Cheng (1983), and others cited by Xiaobai (1984), have explained its formation in terms of an annual flux divergence at the base of the active layer, as outlined above. It is important to note that the Cheng hypothesis implies that rather than there being a "long-term ice balance" in the uppermost permafrost zone (see Mackay 1983a, p. 121), permafrost is a net sink for atmospheric water.

These processes are not only relevant to the hydrology and

hydrogeology of permafrost, but are also significant to the geomorphology of permafrost regions. In particular, Washburn (1983) has suggested that the growth of palsas in the continuous permafrost zone may, in part, be due to downward water movement associated with a rising permafrost table. The Cheng hypothesis suggests that such features may form independently of a rising permafrost table, as long as frost-susceptible sediments underlie the organic surface material. Additionally, the processes may be responsible for the development of frost-heaved bedrock in permafrost regions (Dyke 1979, 1981; Dionne 1983), due to the ability of certain rocks to ~~contain~~ and conduct unfrozen water (Mackay 1984b; Hare 1985; Walder and Hallet 1985). Finally, Mackay (1981, Abstract) noted that most downslope soil movement on Garry Island, N.W.T., occurs in late summer due to "frost creep by thaw of an ice-rich layer at the bottom of the active layer." Egginton and French (1985) have reported observations from Banks Island, N.W.T., of "pluglike" slope movement occurring throughout summer associated with thawing of ice lenses at the base of the active layer.

The remainder of this thesis describes experiments and observations carried out to determine the validity of the Cheng hypothesis.

CHAPTER 3

REGIONAL SETTING, SITE CONDITIONS AND EXPERIMENTAL TECHNIQUES

The experiments reported in this thesis were carried out near Mayo, central Yukon Territory. During the Late-Pleistocene glacial periods, central Yukon formed the eastern limit of Beringia, the unglaciated corner of northwest North America that was intermittently connected to Asia by an isthmus across the Bering Sea (Hopkins 1967). The Mayo area was close to, but within, the ice limit during the Wisconsinan glaciation (Bostock 1966; Hughes et al. 1969; Hughes 1983a). As a result, most of the local sediments are of Holocene age.

The experimental sites considered here are either in Recent alluvial deposits, or in glaciolacustrine sediments deposited in a proglacial lake at the end of the Pleistocene. In order to present the experimental work in an appropriate context, the first part of this chapter comprises a brief review of Late-Quaternary environmental conditions in the study area, with emphasis on the last glaciation and the Holocene. Second, a brief description of the present environment is provided, as an introduction to details of the sites studied and experimental techniques used in this thesis.

3.1 Late-Quaternary glaciations in central Yukon.

In terms of Pleistocene geology, central Yukon can be divided into two regions (Hughes 1972). First, the southern and eastern parts of the Territory were covered by Cordilleran ice moving north and west from the Cassiar and Selwyn Mountains (Figure 3.1) (Bostock 1966). Second, the land north of the Ogilvie mountains and west of the Cordilleran ice lobes remained ice-free (Hughes *et al.* 1969, 1981), although some alpine glaciers formed at higher elevations in the Ogilvie Mountains, west of the Cordilleran glacial limits (Vernon and Hughes 1966), and in the McArthur Group and Wernecke Mountains within the limits of valley glaciation (McTaggart 1960; Hughes 1983b). This region was the largest unglaciated area in Canada (Rutter 1984) and became a glacial refugium for plants and animals throughout much of the Pleistocene (Harrington 1978; Ritchie 1984). Sediments in the Old Crow area provide the best record of Pleistocene environmental changes in Canada (e.g. Hughes 1972; Lichti-Federovich 1973, 1974; Morlan and Matthews 1983).

Bostock (1966) mapped four separate (Cordilleran) glacial drift surfaces in central Yukon and named the associated glacial advances (from earliest to most recent) the Nansen, Klaza, Reid and McConnell Glaciations. The Nansen and Klaza advances are usually considered together as the "oldest" or pre-Reid (pre-Pleistocene) glaciation (Figure 3.2) (e.g. Hughes 1983a; Rutter 1984; Tañocai *et al.* 1985).

Although the timing of the Reid glaciation is not well

— — — Maximum extent of Wisconsin glacialation

..... Maximum extent of older Pleistocene glaciations

Figure 3.1 Late-Quaternary ice sheet limits in northwest Canada and eastern Alaska. (Prest 1984, Map 1584A).

Figure 3.2. Glacial limits in central Yukon Territory.

(Tarqocai et al. 1985, Fig. 30.1).

defined (Hughes 1969; Foscoulos et al. 1977), wood found in tephra cut into Reid outwash has been determined as 42.9 ka old (GSC-524, Lowdon and Blake 1968). Furthermore, bone associated with this tephra (the so-called "Dome Ash" or "Sheep Creek" tephra), has yielded U-Th and U-Pa ages of 78 ka and 73 ka respectively (Hopkins 1982; Schweger and Matthews 1985). With these dates in mind, Hopkins (1982) suggested that the Reid Glaciation should be considered an Early-Wisconsinan advance that was over by 80 ka.

The Reid was followed by an interstadial period, during which a diverse vertebrate fauna flourished throughout Beringia (Harrington and Clulow 1973; Guthrie 1982; Matthews 1982). The ages of fossils from this period range between 81 ± 7.5 ka, for a Mammuthus tusk found in the Old Crow flats (Hopkins 1982), and 25.17 ± 0.63 ka, for a similar bone found in the same region (NMC-1232, Morlan 1984). However, glacial conditions were probably established somewhat earlier than 25.2 ka BP, since Picea glauca (white spruce), P. mariana (black spruce) and Betula papyrifera (paper birch) were exterminated from northern Yukon by 30 ka BP (Matthews 1975; Hopkins et al. 1981). Palynological data from Hanging Lake, northern Yukon, indicate a full glacial vegetation at 33 ka BP (Cwynar and Ritchie 1980; Cwynar 1982).

The McConnell limit (Figure 3.2) is marked by fresh glacial landforms in central Yukon (Tarnocai et al. 1985), which join the features of the Kluane, or Macauley, Glaciation to the southwest (Denton and Stuiver 1967; Rampton 1971). The McConnell is also associated with the Donnelly Glaciation in the northern Alaska

Range (Hamilton and Thorson 1983) and the Walker Lake Glaciation in the Brooks Range (Hamilton 1982).

Usually, the McConnell Glaciation is assigned to the Late Wisconsinan (see Hughes 1969, 1983a; Rutter *et al.* 1978), but no finite ^{14}C -dates associated with the beginning of this advance have been reported. However, the pattern of Cordilleran ice advances in central Yukon follows that of southern Yukon (Hughes 1983a), where the Kluane Glaciation began c. 29.6 ± 0.46 ka BP (GSC-769, Denton and Stuiver 1967). Hughes (1983a) suggests that the date of 23.9 ± 1.14 ka (GSC-2811, Klassen 1978) for an advance at Tom Creek in the Liard Plain indicates Cordilleran ice continued to expand after 23.9 ka BP.

The dispersal of Cordilleran ice into central Yukon was controlled to a great extent by topography (Figure 3.3). Glacial limits mapped by Bostock (1966), Hughes *et al.* (1969) and Hughes (1983a, 1983b) indicate that at least McConnell, and possibly Reid, ice was confined to valley bottoms west of 134°W . Glacial limits close to Mayo, elevation 480 m a.s.l., are at 780 m for the McConnell and 1200 m for the Reid Glaciations (Hughes 1983a). Much of the upland terrain in the region was not covered by McConnell ice, and the absence of erratic rocks from Keno Hill, 52 km northeast of Mayo, suggests that Keno Summit, 1970 m a.s.l., was above the ice throughout the Pleistocene (Bostock 1966, Hughes 1983a).

Grey sand wedges of unweathered soil can be found in moderately developed Luvisols and Brunisols of Reid parent material in

Figure 3.3. Generalised McConnell limit and direction of ice movement near Mayo, central Yukon Territory. (Hughes 1983a).

central Yukon (Tarnocai et al. 1985). These sand wedges, and other involutions in Reid drift, formed during the McConnell under a dry climate, 3 to 4°C cooler than today (Péwé 1975; Rutter et al. 1978; Tarnocai et al. 1985). Many of the sand wedges are capped by up to 50 cm of loess, deposited toward the end of the glacial period (Hopkins 1982).

The end of the McConnell is only roughly defined. A sample of organic detritus collected 2 m beneath 9 m of peat at upper Meadowhead Creek (134°W, 63° 18'N), 58 km southeast of Mayo and 2 km inside the local McConnell limit, was 10.34 ± 0.34 ka old (GSC-3901, O.L.Hughes, personal communication, 1984). However, two samples of older organic matter have been reported from within the limits of alpine glaciation in the Ogilvie Mountains: 12.9 ± 0.15 ka BP (GSC-67, Dyck and Fyles 1964); and 13.74 ± 0.19 ka BP (GSC-515, Lowdon and Blake 1968). The latter age, and a comparable determination of 13.66 ± 0.18 ka (GSC-495, Lowdon et al. 1967) for organic material overlying Macauley till, led Tarnocai et al. (1985) to adopt 14 ka as the date of culmination of the McConnell Glaciation.

The pollen evidence from Hanging Lake supports this chronology. Cwynar (1982) reports two pollen zones from the Late Wisconsinan. The first, HL-1 (33 to 18.45 ka BP), is a Herb Zone, representing the full glacial vegetation of northeastern Beringia. The second, HL-2 (18.45 to 14.6 ka BP), contains an increase in Betula pollen, representing a slight amelioration in environmental conditions, with the development of scrub vegetation

in sheltered areas.

3.2 Holocene conditions in central Yukon.

Although conditions began to ameliorate after 18.45 ka BP, a very large influx of Betula pollen was only noticed in the Hanging Lake record at about 11.5 ka BP (Cwynar 1982). Hence, zone HL-3 (14.6 to 11.1 ka BP) was designated the Betula Zone. Similar evidence from Lateral Pond, Bonnet Plume Basin, and M Lake, 15 km southeast of Inuvik, N.W.T., also indicates a Betula rise between 12 and 11 ka BP (Ritchie 1977; Cwynar and Ritchie 1982). At both of these sites, and at Hanging Lake, Picea pollen was recorded about 1 ka after the Betula rise. Picea was only established in the Mackenzie Delta and interior Alaska by 9.05 ka BP (Hyvarinen and Ritchie 1975; Ritchie 1977; Ager 1983), after migrating down the Mackenzie Valley to the Delta (Hopkins et al. 1981) and into Alaska through the lowland passes of the Tintina Trench (Cwynar 1982). The availability of species complicates climatic reconstruction from palynological data in the immediate post-glacial period (Terasmae 1968; MacDonald 1983).

The McConnell ice sheet retreated rapidly in central Yukon after 14 ka BP (Vernon and Hughes 1966; Tarnocai et al. 1985). Large proglacial lakes formed in front of the receding ice margins in many of the valleys, e.g. those of the South McQuesten, Keno-Ladue and Stewart rivers (Hughes 1983a), but since the shorelines of these lakes are subdued, the lakes were probably short-lived (Vernon and Hughes 1966). Nevertheless, lacustrine

deposits "as much as several hundred feet thick" have been described (Green 1971, p. 7), and although this estimate includes the large amounts of ground ice present in the sediments, the deposits are, indeed, considerable (Hughes 1983a).

The proglacial lake that formed in the Stewart Valley was impounded by ice and moraine 1 km west of Mayo (Figure 3.4), and extended east to Fraser Falls and up Watson Creek (see Figure 3.3) (Hughes 1983a). The silts and clays deposited in this lake have generally been covered with alluvial material by the modern river, but patches of lacustrine sediments are still found at the surface in some parts of the valley. These deposits are ice-rich and contain segregated ice lenses up to 30 cm thick, which formed by water migration to the freezing front when the proglacial lake drained and permafrost aggraded into the saturated sediments (Burn *et al.* 1986). Thermokarst lakes and thaw slumps which have formed in these sediments are distinctive features of the geomorphology of the study area (see Frontispiece).

Although the thermokarst lakes are probably recent features, about 100 years old (Burn 1982), evidence of earlier thermokarst activity is provided by organic beds exposed in the headwall of the largest thaw slump 3 m below the present ground surface. These layers, which contain various plant macrofossils, amongst them Salix, (willow) Betula and Picea trunks, have been interpreted as thaw lake bottoms (French *et al.* 1984). Two ^{14}C ages of 8.56 ± 0.13 ka (BGS-841) and 8.52 ± 0.12 ka (BGS-842) have been determined for wood pieces extracted from the layers (Burn *et al.*

Figure 3.4 McConnell glacial limit and western limit of glaciolacustrine deposits near Mayo, Yukon Territory. (after Hughes 1983a).

1986).

Brown (1973) and Rampton (1973) report that ground ice melting and thermokarst development was common in the Mackenzie Delta after 12 ka BP, and there is considerable evidence that optimal post-glacial climatic conditions were reached in the early Holocene (Delorme et al. 1977; Mackay 1978; Rampton 1982; Ritchie et al. 1983). Reconstructions of mean annual air temperature from ostracod assemblages in lake sediments by Delorme et al. (1977) suggest that between 14.4 and 6.8 ka BP the temperature in northwest Canada rose by up to 11.6°C , to between $+0.8$ and $+1.5^{\circ}\text{C}$. This is, undoubtedly, an overestimate, since such a large increase in air temperature would have led to thawing of massive icy beds along the western arctic coast of Canada (Mackay 1978). A more reasonable estimate is of mean annual air temperatures 1 to 2°C warmer than present values (e.g. -9.8°C at Inuvik, N.W.T., -4.0°C at Mayo, Atmospheric Environment Service 1982) (Hopkins et al. 1981). This is obtained by moderating an estimate of 5°C warmer than present for sites on the Tuktoyaktuk Peninsula early in the Holocene (Ritchie and Hare 1971) by 3 or 4°C due to a lower sea level (Mackay 1963; Rampton 1973; Hill et al. 1985). (A sharp increase in summer temperature with distance from the Beaufort Sea is a feature of the climate in the Mackenzie Delta area (Burns 1973).)

Pollen evidence from the period after arrival of most species now in the region indicates that optimum climatic conditions were reached by 8.5 ka BP in the Mackenzie Delta (Hyvarinen and Ritchie

1975; Ritchie 1977; Ritchie et al. 1983), north Yukon (Cwynar 1982), Old Crow area (Lichti-Federovich 1973, 1974) and interior Alaska (Ager 1983). At that time aquatic plants and animals, e.g. Typha latifolia (cattail), Castor canadensis (beaver), extended their ranges well beyond later Holocene limits (McCulloch and Hopkins 1966; Schweger et al. 1982). The evidence of thermokarst activity in the Mayo region at the same time suggests that the climatic optimum in central Yukon occurred contemporaneously.

In addition to the organic beds in the Mayo headwall exposure, the ground ice stratigraphy contains two principal features (Burn et al. 1986):

- (i) Ice wedges truncated about 1.5 m below the bottom of the active layer (Figure 3.5), and
- (ii) A sharp, continuous break in cryotexture of the segregated ice at the same level as ice wedge truncation (Figure 3.6).

The segregated ice below the discontinuity is of the thick cryotexture mentioned above (ice lenses up to 30 cm x 80 cm in cross-section). Above the discontinuity the ground ice is composed of segregated lenses about 2 cm x 15 cm in cross-section (Burn et al. 1986). Organic material recovered from the sediment at the level of the discontinuity has been dated at 8.87 ± 0.2 ka BP (WAT-1180). Burn et al. (1986) examined the stratigraphy and oxygen-isotopes of the ground ice (Figure 3.7) and the clay mineralogy of the sediments associated with different cryotextures. They concluded that the break in cryotexture represents a thaw unconformity, which truncated ice wedges and melted other

Figure 3.5. Truncated ice wedge in the headwall of the larger thaw slump near Mayo, Yukon Territory. 5 June 1985.

Figure 3.6. Thaw unconformity in the headwall of the larger thaw slump near Mayo, Yukon Territory. 11 October 1985.

Figure 3.7 Stratigraphy and oxygen-isotope ratios of ground ice in headwall of thaw slump near Mayo, Yukon Territory. (for detailed explanation see Burn *et al.* 1986, Fig. 5.)

ground ice during the Holocene climatic optimum. Since the $^{18}\text{O}/^{16}\text{O}$ ratio of the ice above the unconformity is less negative than that of the original proglacial lake water below, it is thought to be younger ice, formed from water precipitated during a warmer climatic interval (Burn et al. 1986; cf. Mackay 1983b). Moreover, this upper-layer ice represents a period of permafrost aggradation over the last 7 ka or so, that, initially at least, accompanied the deterioration in climate from optimum conditions.

The deterioration may have begun by 6 ka BP, and certainly started before 5 ka BP (Hamilton et al. 1983; MacDonald 1983; Ritchie 1984). The transition to cooler conditions is apparent in all of the pollen records mentioned above, most of which contain a single unit for the period 6 ka to present, implying relatively constant conditions since 6 ka BP.

However, the Recent glacial record from the St. Elias Range (Denton and Karlen 1973, 1977) and the Brooks Range (Ellis et al. 1981; Ellis and Calkin 1984) indicates several periods of expansion by cirque and valley glaciers in the last 3 ka. Two principal advances, between 2.9 ka and 2.1 ka and 0.5 ka and 0.1 ka ago have been reported from the St. Elias Range (Denton and Karlen 1977), with the Holocene maximum in the latter period or "Little Ice Age" (Grove 1979; Clague and Rampton 1982). A similar glacial maximum has been documented in the Brooks Range, with rapid glacier retreat after 1870 AD (Ellis and Calkin 1984). Warmer conditions between these advances have been inferred from high tree-lines in southwest Yukon (Rampton 1971), and records of

ice-wedge truncation near Fairbanks, Alaska (Hamilton et al. 1983).

Evidence of climatic fluctuations over the last few centuries (e.g. Jacoby et al. 1985) is not visible in the ground ice stratigraphy of the Mayo site. Some numerical modelling has indicated that the thermokarst lakes which dominate the geomorphology of the study area were probably initiated by a local disturbance (forest fire?), rather than climatic change (Burn 1982). However, relict debris-covered glaciers from the recent Neoglacial can be found in some of the north-facing cirques of the McArthur Group, 40 km south of Mayo (Hughes 1983b).

3.3 The study area.

Mayo is located at the confluence of the Mayo and Stewart Rivers at $135^{\circ} 35' W$ $63^{\circ} 35' N$. The Stewart River valley is 5 km wide, and the hills on each side climb from the valley floor at 480 m to 1200 m.

3.3.1 Climate and permafrost.

The mean annual temperature at Mayo between 1951 and 1981 was $-4.0^{\circ}C$, and mean total precipitation 306.3 mm, of which an average of 121.2 mm (water equivalent) fell as snow (Atmospheric Environment Service, 1982).

In Figure 3.8 mean monthly temperatures and precipitation recorded at Mayo Airport during the study period (August 1983 to October 1985) are compared with the Normal values. In general,

Figure 3.8 Normal and observed mean monthly temperature and total precipitation, August 1983 - October 1985, Mayo Airport.

the climate behaved in a Normal fashion throughout the study, although the first four months of 1984 were warm and above average snowfall was recorded in January and February. The freeze-back period (October to December) 1984 was slightly cooler than Normal, but was followed in January 1985 by the warmest departure from Normal recorded during the study (+19.6°C). Table 4.1 indicates the dates of "significant" changes in the air temperature regime, and it is clear that the annual cycle was duplicated in each year. The first permanent snow of each winter fell in the second week of October and finally melted in the third week of May. Snow accumulation over the 1984/85 winter varied at the study sites between 28 and 63 cm on 1 April 1985 (see Figure 4.1).

Permafrost conditions in the area have been summarized in Chapter One (p. 9), but it is important to add that while mesic valley floors in central Yukon are generally frozen, the dry outwash terraces, which are found in the Stewart Valley west of Mayo, are not generally underlain by permafrost. Ground ice has not been reported in any of the gravel pits used for road construction in central Yukon, for example (O.L. Hughes, personal communication, 1985). Much of the Recent alluvium is permafrost-free.

3.3.2 Vegetation and soils.

Mayo is in region B26b of the Boreal Forest (Rowe 1972), and ecoregion 12 (Pelly River) of the Yukon Territory (Oswald and Senyk 1977). The dominant tree species in undisturbed terrain in

Figure 3.12. Soil materials, visible ice contents and volumetric water contents at experiment sites, Mayo study area.

Table 3.1
 Dates of significance in the air temperature regime,
 Mayo Airport, 1983 - 1985.

Year	First day ^a max. T > 0°C	Last spring frost	First fall frost	First day ^b max. T < 0°C
1983			19 Aug.	24 Oct.
1984	3 Mar.	28 May	12 Aug.	25 Oct.
1985	11 Mar.	22 May	29 Aug.	17 Oct.

a. Date after which maximum daily air temperature remained above 0°C.

b. Date after which maximum daily air temperature remained below 0°C.

Sources: Meteorological data log, Mayo Airport; and Monthly Station Data Records, Mayo, Y.T. (Environment Canada).

the study area is Picea glauca but scattered P. mariana specimens may also be found. Betula papyrifera and Populus balsamifera (balsam poplar) grow with P. glauca on dry soils in disturbed terrain, while in mesic locations, near the river, for instance, stands of Alnus crispa (alder) and Salix alaskensis predominate.

The shrub layer is dominated by Ledum palustre (Labrador tea) and Vaccinium spp. (e.g. low-bush cranberry) in wet areas, with Rosa acicularis (wild rose) and Polygonum alpinum (wild rhubarb) at drier sites. There is a rich wild flora, characteristic of the region, dominated by Epilobium angustifolium (Fireweed) in disturbed areas, and a number of species, e.g. Lupinus arcticus (Arctic lupin), Potentilla spp. (Cinquefoil), Pyrola spp. (Wintergreen), in the forest. The forest floor is usually covered by Hypnum spp. (peat), with patches of Cladonia spp. and Peltigera spp. (lichen).

There has been little profile development in the soil, partly because it is young, and partly because low temperatures and impeded drainage have inhibited soil weathering. However, a Bh horizon 10 to 15 cm thick is sometimes found beneath the peat cover.

The study area contains soils of four principal types (see Appendix 1):

- (i) A silt loam derived from river alluvium and colluvial deposits (25 % sand, 68 % silt, 7 % clay);
- (ii) A silt loam derived from the lacustrine sediment (2 % sand, 73 % silt, 25 % clay), which in places is

- (iii) A silty clay loam (1 % sand, 70 % silt, 29 % clay), or
- (iv) A silty clay (1 % sand, 53.5 % silt, 45 % clay).

Permeability tests on several samples of these soils collected at the study sites indicated that the unfrozen hydraulic conductivity of the silt loams and the silty clay loam is about $2 \times 10^{-8} \text{ m s}^{-1}$; and of the silty clay about 10^{-9} m s^{-1} (see Table 3.2). Dry bulk densities of 1.2 g cm^{-3} for the silt loams and 1.14 g cm^{-3} for the silty clay were determined in the laboratory.

The unfrozen water content curves for each of these soils are presented in Figure 3.9. These curves were determined in the laboratory by Time Domain Reflectometry, with a resolution of $\pm 2.5 \%$ in θ_u (Patterson and Smith 1981). The curve for silty clay was drawn from determinations of θ_u made by D.W. Riseborough (Smith and Riseborough 1985). The slightly lower values presented for this curve compared with those for silty clay loam and silt loam are due, perhaps, to small differences in interpretation of the TDR trace, the resolution of the technique, and a small amount of salt in the pore water (see Table 3.3).

However, pore water salinity was generally low (Table 3.3), and hence osmotic effects complicating soil moisture behaviour during freezing are considered negligible.

The clay-size fraction of the lacustrine silty clay is principally composed of quartz, hydrous mica and chlorite. This composition was determined by X-ray diffraction on 45 samples, all of which provided similar results (Burn *et al.* 1986). An example of the diffraction patterns obtained for the material after three

Table 3.2

Hydraulic conductivity of soils from various sites,
Mayo study area.

Site	Depth (cm)	Soil	K (m s^{-1})
1	50 - 60	Silty clay	2.6×10^{-9}
1	130 - 150	Silty clay	5.5×10^{-10}
2	20 - 30	Silty clay loam	2.1×10^{-8}
2	140 - 150	Silt loam	2.2×10^{-8}
3	170 - 180	Silty clay loam	4.5×10^{-8}
3	180 - 200	Silty clay loam	4.5×10^{-9}
6	130 - 150	Silt loam	8.5×10^{-8}
7	180 - 200	Silt loam	3.8×10^{-8}
8	80 - 90	Loam	2.6×10^{-8}
8	90 - 100	Silt loam	1.7×10^{-8}
8	150 - 170	Sandy loam	1.1×10^{-7}
Tubes ^a		Silt loam	4.6×10^{-8}

a. Value for silt loam in soil analogue tubes

(see p. 110).

Figure 3.9 Unfrozen water content characteristic curves for soils at the study sites.

Table 3.3

Electrical conductivity of soil pore water, Mayo study area.

Site	Depth	Conductivity ($\mu\text{mho cm}^{-1}$)	NaCl equivalent (10^{-3} g l^{-1})
1	0 - 10	0.30	0.21
1	70	0.31	0.21
1	130	0.35	0.24
1	180	0.56	0.39
2	80 - 90	1.75	1.40
2	160 - 170	2.10	1.60
3	70 - 80	2.00	1.50
7	30	0.82	0.65

pretreatments (Mg-saturation at room temperature; K-saturation at room temperature; and K-saturation followed by heating to 540°C) is shown in Figure 3.10. An additional treatment of K-saturation followed by heating to 700°C produced diffraction patterns identical to those obtained after heating to 540°C.

Quartz was recognised by a peak at 4.26 Å (Wilding *et al.* 1977), and hydrous mica by peaks at 10 Å and 5 Å (Fanning and Keramidas 1977). The persistence of the 14 Å peak after heating to 540°C indicated the presence of chlorite (Mitchell 1976). The absence of a peak at 17.7 Å after Mg-saturation indicated that montmorillonite was not present in the samples, and, similarly, the lack of lower order kaolinite peaks at 3.57 Å, 2.34 Å, 2.29 Å, or 1.9 Å in the unheated samples indicated that there was no kaolinite present (Carroll 1970). The samples were presumed to have a negligible vermiculite content when a peak at 9.3 Å did not appear after heating to 700°C (Carroll 1970).

3.4 Site descriptions.

Instrumentation was installed at eight sites in the study area (Figure 3.11). Six of these were located in the glacio-lacustrine material, and two were in Recent alluvial sediments. Soil materials, ice and moisture contents of the sites are indicated in Figure 3.12; photographs of the sites are provided in Figure 3.13; and vegetation data are presented in Table 3.4.

Figure 3.10 X-ray diffraction patterns of clay fraction
of silt loam after three pretreatments.

Figure 3.11. Locations of the experiment sites, Mayo study area

Figure 3.12. Soil materials, visible ice contents and volumetric water contents at experiment sites, Mayo study area.

Figure 3.12. Soil materials, visible ice contents and volumetric water contents at experiment sites, Mayo study area.

Figure 3.12. Soil materials, visible ice contents and volumetric water contents at experiment sites, Mayo study area.

Figure 3.12. Soil materials, visible ice contents and volumetric water contents at experiment sites, Mayo study area.

Site 1.

Site 2.

Figure 3.13. Experiment sites 1 - 8, Mayo study area.

Site 3.

Site 4.

Figure 3.13. Experiment sites 1 - 8, Mayo study area.

Site 5.

Site 6.

Figure 3.13. Experiment sites 1 - 8, Mayo study area.

Site 7.

Site 8.

Figure 3.13. Experiment sites 1 - 8, Mayo study area.

Table 3.4

Percent vegetation cover at sites 1 to 8, Mayo study area.

Site	Trees ^a		Shrubs		Herbs		Moss/lichen		Litter	
	S ^b	V ^c	S	V	S	V	S	V	S	V
1	5	10	10	10	10	40	95	99	1	1
2	50	60	2	20	5	25	-	10	100	100
3	-	2	30	50	100	100	-	-	-	-
4	40	45	1	10	0	10	98	98	2	2
5	2	5	5	5	15	15	85	99	-	1
6	25	30	10	10	25	25	95	98	5	2
7	60	45	1	1	10	10	100	100	-	-
8	70	70	-	-	10	10	-	-	100	100

a. Percent sky obscured by canopy.

b. S. Site conditions refer to an area of 2 m² within which the instrumentation was installed.

c. V. Conditions in the vicinity refer to an area of 100 m² about the site.

Note: Cover determined by quadrat sampling.

Site 1.

Site 1 was located on a silty clay hummock at the edge of a mixed Picea/Salix thicket, about 20 m from an active thermokarst lake. The high θ_u of the soil led to an active hydrological regime during the summer thaw, although rapid freezing prevented much water movement in winter.

The surface was covered completely by Hypnum (Table 3.4), and the local dominant trees, shrubs and herbs were P. glauca, Ledum palustre and Vaccinium vitis-idaea (Mountain cranberry) respectively. The surface vegetation was damaged during equipment installation, exposing the sides of the hummock. This led to some deepening of the active layer during the summers of the study period and rapid refreezing in the fall.

Site 2.

Site 2 was situated in a P. glauca forest, with a dry silty clay loam/silt loam soil. The forest canopy intercepted much precipitation, and, as a result, snow accumulation was low. The upper metre of the soil profile was relatively dry throughout the year, although considerable ground ice lies below it.

The ground surface at the site was almost bare of vegetation, but was covered by humus and decaying wood. Some shrubs were found near the site, particularly Rosa acicularis.

Site 3.

This site was installed in aggrading permafrost, in the

emergent sediments of a thermokarst lake. Permafrost was found between 50 and 250 cm depth in a hole drilled 10 m from the site in 1982. Groundwater below the permafrost was under hydrostatic pressure, and rose up the drill hole to 90 cm. The wet soil conditions, fine grained sediments and a deep snowcover, trapped by the surrounding willow bushes, combined to induce considerable frost action at this site. Interpretation of the hydrology was complicated by frost jacking of the neutron probe access tube and thermistor cable by 30 cm over the study period.

The wet ground surface was covered by grasses, with Salix bushes surrounding the site. There were no trees shading the site.

Site 4.

Site 4 was situated in a mature P. glauca forest, with 10 cm of moss covering silt loam. The site was installed over 30 m away from the headwall of the larger thaw slump, but disappeared into the slump on 1 July 1985. The soil was generally dry, and ground temperatures were the coldest of all sites, partly due to the proximity of the headwall promoting lateral cooling (particularly in winter 1984/85), and partly because of a shallow snowcover.

The ground surface was completely covered by Hypnum, with some herbs, particularly V. vitis-idaea (Table 3.4).

Site 5.

Site 5 was located in a swampy depression, 50 m southwest of

site 4. The neutron probe access tube was not installed at this site in 1983 because of groundwater seepage into the drill hole. The water table periodically rose above the ground surface in June 1985. As at site 3, the lacustrine silt loam soil, deep snowcover and wet conditions combined to produce considerable frost action over the year. The vegetation at the site was damaged inadvertently during equipment installation, when the drilling platform was left covering the site over a 10-day period.

The ground surface was covered by mosses and grasses, with some herbs at the edge of the depression, and a few aquaphilic plants such as Equisetum (Horsetail). The site lay to the south of a mature stand of P. glauca, and was surrounded by Picea (both species) and Salix saplings.

Site 6.

Site 6 was installed near site 4, a further 20 m from the thaw slump headwall. The ground conditions were broadly similar to site 4, although the site was mainly covered by lichens rather than mosses, and the organic horizon was thinner. The dominant tree species was P. glauca.

Site 7.

This site was situated 1.5 km west (down river) of the other sites in Recent alluvial silt loam. Surface conditions, however, were similar to site 4. A thick peat mat retarded summer thawing, and led to the thinnest active layer of all sites. Access to the

site was treacherous during freeze-up of the Stewart River, and hence the observation record is less complete.

The surface was completely covered by Hypnum, with a few herbs, dominated by Linnaea borealis (Twinline). The site was open to the southwest, but otherwise well shaded by P. glauca.

Site 8.

This site was located in a grove of Betula and Picea, near a recently (1952) drained lake. The soil was a coarse colluvial material. Permafrost was not present in the surface 2 m, but 18 m away, in the centre of the trees, frozen ground was encountered at 3.25 m, perhaps indicating permafrost aggrading into the former lake talik.

The ground surface was covered by leaf litter and a few grasses. In summer the site was almost entirely shaded by the deciduous trees, but in winter the canopy covered less than 15 percent of the sky. The exposure accelerated spring snow metamorphism and compaction relative to sites in the coniferous forest.

3.5 Instrumentation.

The hydrological and thermal regimes of the surface layers of the ground were investigated at the eight sites with a variety of instruments and techniques. The instrumentation was installed at each site in July 1983 using a Geological Survey of Canada permafrost coring kit. A plywood drilling platform was used to

limit surface damage at each site. The holes were backfilled with drill scratchings.

3.5.1 Thermistors.

Twelve calibrated thermistors ($\pm 0.05^{\circ}\text{C}$, YSI-44004) were placed on a cable and installed at each site at depths of 200, 175, 150, 130, 115, 100, 85, 70, 55, 40, 25 and 10 cm. The cable was attached to a dowel to maintain constant spacing between thermistors. The thermistors were read with a Keithley 135 multimeter, which provided reliable readings, except after two hours operation at -40°C or if wet.

The upper 15 cm of the cable was cased in 2.54 cm ABS pipe, which was painted white in 1984 after early snowmelt around the (black) pipe was observed in April 1984. The ground surface level (0 cm) was marked on the cable and inspected in August each year to determine the amount of frost jacking that had occurred. The frost table was located each week throughout the summer by probing with a steel rod. The depth to frost determined by the steel probe was compared with the position of the 0°C isotherm derived from the thermistor records. If these values differed by more than 5 cm, and a similar amount of jacking was noticed on the cable, then it was assumed that each thermistor had been moved upward in the soil profile. The calculated temperature field was then realigned using these data.

Over the study period, one to three thermistors per cable apparently drifted from the initial calibration. Temperatures were

interpolated for these points from the adjacent thermistors, except at the 200 cm level.

3.5.2 Lysimeters.

A lysimeter, made from a can of radius 4.9 cm and depth 13.4 cm was installed at each site in mid-June 1984, in order to estimate precipitation and evaporation during the summer and to compare the indices obtained for these variables at different sites. Riseborough (1985) has indicated some of the problems associated with the use of lysimeters to measure evaporation, particularly the discontinuity introduced at the bottom of the lysimeter in the path of upward water movement. However, this effect is minimized in organic soil due to poorly developed hydrologic pathways (Proctor 1982). During work in the study area in 1983, Riseborough (1985) found that lysimeters similar to those used in this research produced reliable measurements of potential evaporation under wet conditions, and adequate representations of evaporation as the organic matter dried. In the light of Riseborough's (1985) observations, these instruments are regarded as adequate for this study.

The lysimeters were placed in a plastic sleeve in order to prevent debris accumulation on the sides of the can, and to facilitate removal of the can from the ground without disruption to the surrounding soil. The surface of the lysimeter was always within 0.5 cm of the soil surface. The lysimeter was weighed at least once every 48 hours during the summer with a portable

balance (Sartorius 1006MP9) to ± 1 g (± 0.2 mm water), less than 0.5 percent of the weight of any lysimeter. The lysimeters were unreliable when the watertable rose above the can, a phenomenon encountered at site 5.

Precipitation at each site was determined from lysimeter measurements taken after rainfall. However, since the lysimeters were not weighed immediately after each rainstorm, the measurements also include evaporation that occurred after the rainstorm and before measurement. Hence, the measurements are of "net precipitation". The reverse applies to measurements of evaporation, and these are reported as "net evaporation".

Any free water in the cans was drained after each measurement, in order to minimize the difference in water content between the can and surrounding soil.

3.5.3 Neutron probe.

Moisture content profiles of the surface 2 m were determined by neutron moderation, with probe model CPN-503. The probe was lowered into the soil via a 6.3 cm ID aluminum pipe.

The probe operates by emitting high energy neutrons and counting the number that return at low speed after colliding with hydrogen atoms. For a given soil layer, the hydrogen content depends on the moisture and organic matter contents, but changes over the time period considered here may be regarded as due purely to changes in moisture content.

Since the instrument records changes in water content by

signal moderation, each observation is generated by a soil volume of indefinite size. Manufacturer's specifications indicate that 90 percent of each signal is derived from a cylinder of radius 15 cm and length 15 cm in most soils, which declines in radius to 7.5 cm for pure water. However, the received signal is directly proportional to the water content and thus if the soil is horizontally homogeneous, the variation in field of view should not lead to complications in monitoring changes in the moisture content profile.

The moisture content of a soil layer is obtained by dividing the count for the soil layer by a "standard" count, taken with the probe in its casing. This ratio (r) is directly proportional to the hydrogen content of the soil layer. The instrument was calibrated for most soil layers by comparing readings taken immediately after the access pipe had been installed with the moisture content of core samples determined gravimetrically. The manufacturer's calibration of $0.45 \Delta r = \Delta \theta_v$ ($\text{cm}^3 \text{ cm}^{-3}$) produced satisfactory determinations of changes in the volumetric water content of individual soil layers over time, as it described the slope of the relationship between water content and r for core samples of similar soil type.

Drift in the performance of the instrument may be detected by examining the standard counts collected over some time. For this study, 19 standard counts taken between 1 June and 2 August 1984 and 19 others between 31 May to 1 August 1985 were compared, using Student's t -test (Ebdon 1985). No significant difference (i.e.

drift) was detected between the samples at the 0.05 significance level (see Table 3.5). The observed difference in means was 0.5 percent.

In addition, the resolution of the instrument was established by repeated counts over two profiles. Table 3.6 displays the range in r after 10 counts over several intervals at each site. The range about each mean r varied between 0 and 2.6 percent (mean range 0.9 percent, standard deviation 0.66 percent). The upper 95 percent confidence limit for the resolution is, therefore, 2.19 percent, and for practical purposes the resolution is ± 1 percent in r , or $\pm 0.005 \text{ cm}^3 \text{ cm}^{-3}$ in water content ($\pm 0.07 \text{ cm}$ over a 15 cm interval).

The neutron probe has generally been used to monitor the moisture content of unfrozen soils (e.g. Shirazi and Isobe 1976). However, since the present study of soil moisture covers a range of temperatures and different phases, it was important to determine whether the neutron probe reading exhibits any systematic temperature-dependency. Studies of the soil-moisture regime during freezing known to the author have not considered such an effect (e.g. Wright 1981; Cheng 1982; Rouse 1984; Hayhoe and Bailey 1985).

Laboratory tests with the neutron probe were carried out at Carleton University (see Appendix 2). The results of the tests are summarized in Table 3.7a, and indicate that no temperature-dependency per se was detected in the performance of the probe. r remained constant for dry soil between $+17$ and -26°C ; for wet soil

Table 3.5

Calculation of Student's t for neutron probe standard counts.

	1984 (x)	1985 (y)
Observations	19 (n_x)	19 (n_y)
Mean	24370 ($\bar{x}$)	24537 ($\bar{y}$)
Sum of squares	5.9395×10^8 (Σx^2)	6.021×10^8 (Σy^2)
Means squared	5.9392×10^8 ($\bar{x}^2$)	6.0208×10^8 ($\bar{y}^2$)

$$t = \frac{|\bar{x} - \bar{y}|}{\sqrt{\left(\frac{\Sigma x^2/n_x - \bar{x}^2}{n_x - 1} + \frac{\Sigma y^2/n_y - \bar{y}^2}{n_y - 1} \right)}}$$

$$t = 1.77$$

with $(n_x + n_y - 2 =)$ 36 degrees of freedom, critical t at 0.05 significance is 2.02. Therefore, H_0 , the samples are drawn from one population, cannot be rejected.

Table 3.6

Range in r from 10 counts over each of several intervals
at two sites, Mayo study area.

Depth	Site A ^a			Site B ^b		
	Max.	Min.	Range (% mean)	Max.	Min.	Range (% mean)
0 - 15	1.42	1.39	2	1.8	1.8	0
15 - 30	2.22	2.19	1.3	1.66	1.65	0.6
30 - 45	2.29	2.27	0.9	1.62	1.61	0.6
45 - 60	2.32	2.30	0.9	1.64	1.62	1.2
60 - 75	2.42	2.41	0.5	1.69	1.68	0.6
75 - 90	2.09	2.08	0.5	1.76	1.76	0
90 - 105	2.05	2.03	1.0	1.73	1.70	1.7
105 - 120	1.91	1.89	1.0	2.03	2.01	1.0
120 - 135	1.86	1.84	1.1	2.18	2.16	1.0
135 - 150	1.83	1.82	0.5			
150 - 165	1.91	1.86	2.6			
165 - 180	1.97	1.96	0.5			
175 - 190	2.03	2.01	1.0			
183 - 198	2.24	2.21	1.3			

a. Data collected at site 3, 8 June 1984.

b. Data collected at a site near Mayo Airport, 26 April 1985.

Table 3.7a

Laboratory determinations of the variation in r with temperature.

Material	Temperature ($^{\circ}\text{C}$)	r
Ice	-26.0	2.03
Ice	-0.3	2.03
Dry sand	20.0	0.02
Dry sand	-26.0	0.02
Saturated silt loam	20.0	1.12
Saturated silt loam	1.5	1.12

Table 3.7b

Laboratory observations of changes in r associated with freezing and thawing.

Material	Temperature ($^{\circ}\text{C}$)	r
Ice	-0.3	2.03
Water	20.0	2.18
Saturated sand	-3.0	0.76
Saturated sand	partially thawed	0.70
Saturated sand	20.0	0.69
Saturated silt loam	-10.0	0.92
Saturated silt loam	20.0	1.00
Silt loam, 90 % saturation	20.0	0.87
	-10.0	0.79
	20.0	0.84
2nd freeze	-10.0	0.74
	20.0	0.82
3rd freeze	-10.0	0.75
	20.0	0.83

between $+20^{\circ}\text{C}$ and $+1.5^{\circ}\text{C}$, and for ice from 0 to -26°C .

However, changes in r were noticed during the melting of ice and thawing of soil (see Table 3.7b). First, r increased by 9 percent between readings taken in ice and water. A similar increase was observed over several readings taken in a melting mixture of crushed ice and water. This is due to the increase in hydrogen density that occurs as ice melts. Second, a 9 percent decrease in r was recorded between counts taken in frozen and thawed saturated sand. A median value was determined from the sand before it thawed completely. Third, a 9 percent increase in r was recorded as saturated silt loam thawed.

Wright (1981, 1983), Rouse (1984), and Smith (1985a) have used the technique to monitor changes in the moisture content profile of the active layer and adjacent permafrost. It is interesting that these authors recorded an apparently sharp increase in the moisture content of soil layers as the freezing front passed through each layer, which was not accompanied by an equivalent desiccation lower in the profile (see Wright 1981, Fig. 6; Rouse 1984, Fig 1; Smith 1985a, Fig. 7). Both Rouse and Smith (personal communications 1984) have indicated that there was no snowmelt or surface infiltration in the period. The neutron probe apparently overestimates the water content of frozen soil. Other authors, e.g. Hayhoe and Bailey (1985), report similar behaviour of the neutron probe moisture-content determination during seasonal freezing, but none have considered this unusual.

These field observations and the laboratory tests with sand

indicate that the neutron probe may increase the estimate of volumetric water content during soil freezing. This is a paradoxical effect, since the number of hydrogen atoms per unit volume of soil may decline as water freezes. The test results of silt loam do not conform to this pattern, but rather to those obtained with bulk water. It is clear that the behaviour of the neutron probe signal during soil freezing and thawing is not fully understood for some soils. This is unfortunate, since the neutron probe is the only instrument that purports to measure the in situ total water content of soil profiles in a practical manner. It is clear that the problem remains to be solved.

Field results similar to those reported above were obtained at Mayo. The readings indicated an increase in the total water content of the active layer during freeze-back, and a decrease during the summer. However, there was no snowmelt during soil freezing and in order to interpret the readings in a physically rational manner, a purely pragmatic approach was adopted. The apparent increase in each layer was corrected by the mean apparent increase in water content of the active layer during freezing. If this was x , and the increase in layer i x_i , the "real" water content increment of layer i after freezing was estimated as $x_i - x$.

For instance, consider the freezing of a four-layered soil of uniform water content $0.4 \text{ cm}^3 \text{ cm}^{-3}$. Suppose that during soil freezing, the apparent water content of the first layer increased by $0.1 \text{ cm}^3 \text{ cm}^{-3}$, that of the second by $0.08 \text{ cm}^3 \text{ cm}^{-3}$, the third by $0.06 \text{ cm}^3 \text{ cm}^{-3}$ and the fourth by $0.04 \text{ cm}^3 \text{ cm}^{-3}$. The mean increment

is $0.07 \text{ cm}^3 \text{ cm}^{-3}$. Substituting the appropriate values for the first layer, the estimate of the water content of this layer after freezing is:

$$\begin{aligned} & 0.4 + 0.1 - 0.07 \text{ cm}^3 \text{ cm}^{-3} \\ & = 0.43 \text{ cm}^3 \text{ cm}^{-3} \end{aligned}$$

and of 0.41 , 0.39 and $0.37 \text{ cm}^3 \text{ cm}^{-3}$ for the second, third and fourth layers respectively. Without a correction of this nature, moisture content increases would be overestimated during freezer-back and underestimated in summer.

3.5.4 Incremental soil displacement tubes.

A set of incremental soil displacement tubes, designed originally by J.R. Mackay (Mackay *et al.* 1979; Mackay 1983a) was installed at each site. The sets consisted of 7 concentric aluminum tubes, whose lower ends terminated at depths of 25, 50, 75, 100, 125, 150 and 200 cm. (see Mackay 1983a, Fig.2). At sites 1 to 7, the tubes which reached 200, 150 and 125 cm were frozen in permafrost. The upper ends of the tubes were separated by 5 cm initially, and as the soil heaved or settled over the period of observation, displacement in the various layers of soil was recorded by changes in separation of the upper ends of the tubes (Figure 3.14). Both Mackay (1983a) and Smith (1985a) have used this technique to monitor soil heave over various time periods.

The location of measured heave is not precisely defined with this apparatus, for an adfreeze bond must develop between the tube and frozen soil before it will move. If frost heave occurs in a

Figure 3.14 Schematic diagram of the telescoping heave tubes.

Note:

Consider the case of soil thickness y cm being increased to $(y+e)$ cm as a result of frost heave.

If an adfreeze bond has developed between the soil and tube 2, in the region AB, then tube 2 will move relative to tube 3, and distance DE will decrease e cm. If the adfreeze bond has not developed, but the soil is bonded to tube 1, then DC will shorten by e cm.

particular layer before bonding, only the tube in the layer above will register the displacement (see Figure 3.14 and accompanying note). Further, measurement of secondary heave and heave due to downward water migration involves tube displacement relative to frozen soil. In these cases, the adfreeze-heave bond must overcome adfreeze-frictional bonds between the tube and stationary soil, it is not clear how much heave need occur before the tubes are displaced. However, since adfreeze bonds are strong, and lead to jacking of piles buried deep in permafrost (Johnston 1981) it is likely that only a small heave is required before displacement is recorded.

An operational problem arose with this apparatus when mineral grains were jammed between the tubes during summer settlement. As a result the tubes revealed only total heave of the active layer with respect to permafrost during the second year of observations.

3.5.5 Soil displacement magnets.

In July 1984 ring magnets, of outer radius 39.2 mm, inner radius 14.2 mm and thickness 4.75 mm, encircling a teflon tube, were buried at intervals of between 10 and 25 cm to depths of up to 1.6 m at 5 sites. Each magnet was located with an accuracy of ± 0.5 mm, relative to one at the soil surface, by a reed switch attached to a graduated cable. Over the summer and in October 1984 the location of each magnet was determined with an apparatus designed to provide readings to within 0.05 mm. This device proved unreliable, and was substituted with the system described.

Measurements of magnet displacement were only satisfactory from November 1984 on.

The magnets moved as the soil heaved or consolidated, but did not provide the interpretation problems presented by the heave tubes, because of their discrete location within the soil. A similar apparatus, the Magnetic Extensometer System, is available commercially from Geotechnical Instruments Ltd., Leamington Spa, England.

Since overall heave detected in the second winter by the heave tubes was within a few millimetres of heave recorded by the magnets, both systems are regarded as reliable indicators of soil displacement (Table 3.8).

3.5.6 Soil analogue tubes.

At most of the study sites, two 10 cm ID x 2 m ABS pipes were fitted into drill holes and filled with six 2.5 cm OD transparent PVC (Tygon) tubes. These tubes were filled incrementally with a silt loam slurry, which was allowed to freeze before a further layer was added. The silt loam was produced in the field from a sandy loam by slurring, sieving and settling to remove any gravel and the bulk of the coarse sand. The soil was placed in each tube by hosepipe, with care taken to avoid trapping air pockets in the slurry. The soil moisture content was above saturation, and at the beginning of July 1984 excess moisture was removed and replaced with soil.

The tubes act as closed-system analogues of the active layer

Table 3.8

Total displacement (cm) recorded by heave tubes and magnets,
8 November 1984 - 29 April 1985.

Site	7/11/84 - 29/1/85		29/1/85 - 29/4/85	
	Magnets	Tubes	Magnets	Tubes
1	0.5	0.13	0	0
3	5.4	6.09	2.6	2.47
5	6.0	5.03	0.1	0
6	0.85	0.89	0.2	0
9	1.9	1.96	2.45	2.51

and surface permafrost layer. At intervals, a tube was extracted and cut in 10 cm sections, which were analysed, gravimetrically, for water content. The growth of ice lenses in the tubes was also observed, particularly just below the base of the active layer.

3.6 Observation schedule.

The instrumentation installed at each site is summarized in Table 3.9. Observations at most sites were made twice weekly during the summer field seasons (1 June to 30 August 1984; 31 May to 1 August 1985), and in November 1984 and April 1985. Readings were taken once a week in September and October 1984, and each fortnight during the winter 1984/85. In addition, three visits were made to the sites over fall and winter 1983/84: in mid-September, early January and early April. The monitoring program was initiated in the first two weeks of August 1983, and the final data were collected in early October 1985.

Most measurements were made by the writer (or field assistants). During September, late October, and early November 1984, and from February to mid-April 1985, readings were taken by Michael Mason-Wood of Mayo.

Table 3.9

Instrumentation installed at each site.^a

Site	Thermistors	Neutron Probe	Heave Tubes	Magnets	Lvsimeter	Analogue Tubes
1	x	x	x	x ^b	x	x
2	x	x	x		x	x
3	x	x	x	x ^b	x	x
4 ^c	x	x	x		x	x
5	x	x ^d	x	x ^e	x	x
6 ^g	x	x	x	x ^e	x	
7	x	x	x		x	x
8	x	x	x	x ^h	x	

a. All instrumentation was operational from 1 August 1983 to 8 October 1985, except:

b. Installed 22 July 1984.	c. Site destroyed 1 July 1985.
d. Installed 20 July 1984.	e. Installed 27 August 1984.
f. Installed 2 July 1984.	g. Installed 16 August 1983.
h. Installed 24 July 1984.	

CHAPTER 4

RESULTS AND DISCUSSION

The site descriptions in the previous chapter have indicated the variety of surface and subsurface conditions found in the study area. These micro-environments are reflected in the diversity of soil thermal and hydrological regimes encountered at the sites, which are now reported. First, the thermal regimes of the study sites are described, since the hydrology of frozen ground depends upon the thermal regime. Second, the hydrologic linkages between the organic surface layer and mineral soil beneath are considered, and detailed analyses of seasonal moisture regimes at particular sites presented. Third, the annual water balance at each site is summarized, and the movement of water into and out of permafrost on an annual basis is discussed. Finally, evidence regarding the Cheng hypothesis for the growth of aggradational ice is summarized and evaluated.

4.1 Soil thermal regime.

4.1.1 Annual temperature cycle.

Geotherm plots describing the soil thermal regime at each site for the year beginning 1 June 1984 are presented in Figure 4.1. The thermal regime in this period was representative of ground temperatures over the two years of the study.

The variation in thermal regimes is indicated in Figure 4.1

Figure 4.1. Soil thermal regime, 1 June 1984 to 1 June 1985,
at sites 1 - 8, Mayo study area.

Figure 4.1. Soil thermal regime, 1 June 1984 to 1 June 1985,
at sites 1 - 8, Mayo study area.

Figure 4.1. Soil thermal regime, 1 June 1984 to 1 June 1985,
at sites 1 - 8, Mayo study area.

Figure 4.1. Soil thermal regime, 1 June 1984 to 1 June 1985,
at sites 1 - 8, Mayo study area.

by differences in the pattern of active layer development at each site. The active layer was shallowest at site 7 (72 cm) and deepest at site 1 (110 cm), with a mean depth of 90 cm at the seven sites with permafrost. During September the active layer cooled rapidly at all sites and soil freezing began by mid-October. Freezing was slowest at sites 3 and 5, where the soils were wettest, but a characteristic feature at all sites was the slow freezing of pore water below 50 cm, which extended the -0.5°C geotherm into December or January. The slower cooling rate coincided with snow accumulation at the sites.

The warmest winter permafrost conditions were observed at site 3, due to the proximity of the lake talik below, the deep snowcover, and the high soil moisture content. The coldest ground temperatures were recorded at site 4 because of a thin snowcover, due to the site's exposed location, and the proximity (5 m) of the exposed thaw-slump headwall, which promoted lateral cooling. Coldest surface conditions were reached at all sites by early March, but rapid ground warming did not start until mid-April, when the refreezing of snowmelt infiltration contributed somewhat to the ground heat flux. The ground was still frozen when the last observations of the winter were made on 2 May 1985, but had begun to thaw by the end of May.

At site 8, where there is no permafrost, the seasonal frost penetrated the soil rapidly during the first three weeks of the freezing period, but slowed after some snow accumulated. The dry

soil thawed rapidly at this site.

Mean annual ground temperatures and the maximum active layer depth at each site are presented in Table 4.1. A "thermal offset" is observed at all sites, with mean annual temperatures in the active layer higher than in permafrost. The offset is due to the higher thermal conductivity of soil when frozen than when thawed (Goodrich 1978), and the moderating influence of snowcover on ground temperatures in winter (Smith 1975; Goodrich 1982). Similar near-surface mean annual ground temperature profiles have been reported by Goodrich (1983, Figs. 9-11) from a section of the Mackenzie Highway near Wrigley, N.W.T.

The data in Table 4.1 indicate that the offset terminates a short distance below the base of the active layer, where there is little seasonal change in thermal conductivity (Goodrich 1978). During the winter, ground temperatures at site 4 were lower than normal due to the proximity of the thaw-slump headwall, and hence the offset does not terminate at this site. At site 3 the offset continues into the permafrost, since the temperature range, e.g. -0.65 to -0.08°C at 85 cm, leads to considerable seasonal variation in soil thermal properties (e.g. see Kay et al. 1981).

The data from sites 3, 5 and 7 support Goodrich's (1978, Fig. 3) demonstration that permafrost may be preserved in environments where the mean annual surface temperature is above 0°C . The mean annual temperature gradient below the offset at site 8 is $2.7^{\circ}\text{C m}^{-1}$, two orders-of-magnitude greater than the geothermal gradient in the region ($0.023^{\circ}\text{C m}^{-1}$, see p. 39). This suggests

Table 4.1

Mean annual temperature profiles at the study sites.

Depth (cm)	Site							
	1	2	3	4	5	6	7	8
10	-1.36	-0.81	0.62	-1.74	0.32	-0.35	0.14	1.93
25	-1.48	-0.88	0.57	-2.27	-0.09	-0.94	-0.76	1.23
40	-1.56	-1.00	0.18	-2.53	-0.51	-1.43	-1.19	0.86
55	-1.72	-1.13	0.00	-2.66	-0.74	-1.46	-1.39	0.43
70	-1.96	-1.27	-0.09	-2.77	-0.96	-1.31	-1.47	0.35
85	-2.11	-1.20	-0.26	-2.78	-1.18	-1.35	-1.48	0.51
100	-2.20	-1.20	-0.40	-2.84	-1.36	-1.30	-1.46	0.86
115	-2.16	-1.18	-0.49	-2.88	-1.42	-1.38	-1.53	1.31
130	-2.14	-1.25	-0.46	-2.94	-1.48	-1.48	-1.48	1.77
150	-2.17	-1.24	-0.39	-3.00	-1.45	-1.46	-1.51	2.32
175	-2.16	-1.24	-	-2.93	-1.40	-1.60	-1.48	2.91
200	-	-1.13	-	-3.06	-1.37	-	-1.47	3.46

Maximum active-layer depth (cm)

110

90

80

89

95

90

72

Temperatures in °C.

that the warm temperatures are a residual feature of the former lake talik.

4.1.2 Temperature gradients.

The ground temperature gradient is a critical factor governing water movement in frozen soil. At all sites a negative temperature gradient, i.e. temperature declining with depth, was observed in the surface layers just before the start of snowmelt infiltration. The negative gradient was recorded for the first time in 1985 on 22 April, and snowmelt began shortly thereafter. The magnitude of the gradient was on the order of 2 to 3°C m⁻¹ at all sites during snowmelt, and at the base of the active layer in June and July. It declined in August and September as the penetration of the thawing front slowed.

In 1984, freeze-back began by 15 October at all sites, and terminated by 20 December, except at sites 3 and 5. A monotonic temperature gradient was only established at site 3 by 1 April and at site 5 by 12 January 1985. However, ground temperatures were generally below 0°C by 15 November, sites 3 (by 16 February 1985) and 5 (by 26 November) being the exceptions again. Figure 4.1 indicates that there was relatively little upward freezing from the base of the active layer during freeze-back, as noted by Parmuzina (1978) and Smith (1985a). Upward freezing was greatest at the dry sites (2, 4 and 7), where there was least latent heat to dissipate.

Positive temperature gradients on the order of 2 to 4 °C m⁻¹

were established throughout the profile during freeze-back. Such values were observed continually over the winter in the lower parts of the profile, but steeper gradients were observed close to the ground surface. A maximum value of $9.3^{\circ}\text{C m}^{-1}$ was measured between 10 and 25 cm on 2 March at site 4, where the snowcover was thinnest.

Overall, the change in the modulus of the temperature gradient between summer and winter at the base of the active layer was small, on the order of 1 to 2°C m^{-1} , or a factor of 2. However, the accompanying change in temperature was adequate to lower the hydraulic conductivity by at least an order-of-magnitude. Whilst the direction of water movement is determined by the temperature gradient, net flow is governed by the temperature-dependent hydrological properties.

4.2 Soil hydrological regime.

4.2.1 Boundary layer.

The soil thermal and hydrological regimes are, of course, linked to the atmosphere. However, the atmosphere's influence on soil conditions is moderated by the processes in the boundary layer of vegetation and surface organic material separating it from the mineral soil (Luthin and Guymon 1974). In addition, there is a snowcover in winter. This "buffer layer" is relevant to the soil thermal regime throughout the year (Smith 1975; Smith and Riseborough 1983), but it exerts less direct influence on the soil hydrological regime in winter than during snowmelt and in

summer, when soil moisture recharge and the bulk of evaporation occur (Rouse 1982b; Kane and Stein 1983b).

Riseborough (1985) has shown that in thermal terms, the most important part of the buffer layer in summer is the surface peat cover. The peat mosses in the Mayo area are Hypnum spp. (identified by Dr. I.L. Bayly, Department of Biology, Carleton University); these are ectohydric bryophytes, which function on water stored after the plant's surfaces are moistened by rain or dew (Proctor 1982). These species absorb and lose water over their entire surface area, and operate in effective isolation from the mineral soil. Attachment to the ground is to procure stability rather than moisture, and often a 5 to 10 cm gap between moss and mineral soil can be found around the edges of hummocks in the study area. In contrast, most vascular plants are endohydric and derive their moisture from the mineral soil via a root system.

The importance of these bryophytes to the hydrology of the study area is that an unsaturated surface organic layer prevents downward water movement into mineral soil. Indeed, after examining the moisture-retention properties of organic soil in central Alaska, Santeford (1979, cited in Chacho and Bredthauer 1983) reported that drainage through the organic horizon only occurs after it is saturated. Otherwise, precipitation is absorbed by the moss and utilized in respiration. Therefore, when the organic layer is dry, the mineral soil is partly detached from the atmosphere: recharge by precipitation is severely restricted, but desiccation through evapotranspiration continues.

Lysimeter measurements of precipitation and evaporation at each site are summarized in Table 4.2. The lysimeters provide a useful index of the relative amounts of precipitation received at the various sites. At most sites, net precipitation was less than that measured at Mayo Airport due to interception by the canopy and evaporation from the lysimeter that occurred after rainfall and before measurement (see p. 99). However, an inverse relationship between canopy closure and measured precipitation is detected from inspection of Tables 3.4 (p.92) and 4.2, as reported by Riseborough (1985). The data for site 5 in June 1985 are complicated due to the water table being above the ground surface from time to time.

The net "evaporation" at each site is also indicated in Table 4.2. While these results must be treated with caution (see p. 98), they confirm that when precipitation is inadequate to saturate the surface (e.g. as in July 1985 when no water was drained from the lysimeters), net evaporation is approximately equal to net precipitation. This suggests that for much of the summer total evaporation is approximately equal to total precipitation, and that the surface is hydrologically independent of the mineral soil below (cf. Wright 1981).

A comparison of net moisture content changes in the surface layer (as measured by the lysimeters), and of soil moisture changes in mineral soil (detected by the neutron probe), is presented in Tables 4.3a and 4.3b. The water content of the mineral soil did not increase over any of the periods described,

Table 4.2

Net precipitation and evaporation measured by lysimeters
at sites 1 - 8, July 1984, and June and July 1985.

Net precipitation (cm).

	Site							
	1	2	3	4	5	6	7	8
July 84	1.47	2.24	2.29	0.69	1.03	1.16	2.76	2.19
June 85	2.69	3.44	3.80	2.33	6.97	3.44	3.41	3.40
July 85	2.40	1.19	1.72	-	3.75	2.95	0.78	1.42

Precipitation recorded at Mayo Airport:

July 84 3.39 cm; June 85 4.71 cm; July 85 6.52 cm.

Net evaporation (cm).

	Site							
	1	2	3	4	5	6	7	8
July 84	1.45	1.73	1.21	1.15	4.01	2.81	1.21	2.14
June 85	2.05	1.60	1.64	2.23	2.11	2.03	1.31	1.62
July 85	1.56	1.40	1.29	-	2.36	2.00	0.84	1.55

Net change (cm).

	Site							
	1	2	3	4	5	6	7	8
July 84	+0.02	+0.51	+1.08	-0.46	-2.98	-1.65	+1.55	+0.05
June 85	+0.65	+1.84	+2.16	+0.09	+4.86	+1.41	+2.10	+1.78
July 85	+0.84	-0.21	+0.43	-	+1.39	+0.95	-0.06	-0.13

Table 4.3a

Net change in water content (cm) of organic and mineral horizons.

July 1984 and June and July 1985.

	Site 1		Site 2		Site 3		Site 4	
	L ^a	NP ^a	L	NP	L	NP	L	NP
July 84	+0.02	-1.6	+0.51	-7.6	+1.08	-2.9	-0.46	-4.3
June 85	+0.65	-3.2	+1.84	-8.9	+2.16	-5.5	+0.09	-1.5
July 85	+0.84	-3.0	-0.21	-0.8	+0.43	-1.8	-	-

	Site 5		Site 6		Site 7		Site 8	
	L	NP	L	NP	L	NP	L	NP
July 84	-2.98	-	-1.65	-10.4	+1.55	-0.5	+0.05	-4.9
June 85	+4.86	0	+1.41	-0.7	+2.10	-4.2	+1.78	-4.0
July 85	+1.39	0	+0.95	-2.0	-0.06	-2.2	-0.13	-1.4

a. L : Lysimeter.

NP: Neutron probe measurement of soil desiccation.

Table 4.3b

Net lysimeter evaporation and mineral soil desiccation (cm),
July 1984 and June and July 1985.

	Site 1		Site 2		Site 3		Site 4	
	L	NP	L	NP	L	NP	L	NP
July 84	1.45	1.6	1.73	7.6	1.21	2.9	1.15	4.3
June 85	2.05	3.2	1.60	8.9	1.64	5.5	2.23	1.5
July 85	1.56	3.0	1.40	0.8	1.29	1.8	-	-

	Site 5		Site 6		Site 7		Site 8	
	L	NP	L	NP	L	NP	L	NP
July 84	4.01	-	2.81	10.4	1.21	0.5	2.14	4.9
June 85	2.11	0	2.03	0.7	1.31	4.2	1.62	4.0
July 85	2.36	0	2.00	2.0	0.84	2.2	1.55	1.4

a. L : Lysimeter.

NP: Neutron probe measurement of soil desiccation.

due to evapotranspiration by trees, shrubs and herbs (cf. Wright 1981; Rouse 1984). In contrast, the moisture content of the surface layer either increased, or declined by a relatively small amount.

Several papers reviewed by Rutter (1975) indicate that the ground surface accounts for only 10 to 12 percent of total evaporation from coniferous forests. The data in Table 4.3a are not directly comparable with these figures, since they do not account for the evaporation of precipitation intercepted by vegetation. In addition, it is difficult to isolate inflow and outflow (evapotranspiration) from the mineral soil, when the rate of infiltration is small because of low soil permeability. As a result, changes in the water content of mineral soil measured by the neutron probe are underestimates of evapotranspiration. However, Table 4.3a does indicate that, in hydrologic terms, the two layers function independently.

In Table 4.3b, the same neutron probe data are compared with net evaporation as determined by the lysimeters. Given that the neutron probe observations underestimate evapotranspiration, these data compare favourably with the data presented by Rutter (1975), except at site 5, where no net soil moisture loss was measured, as the soil remained saturated.

In summary, the lysimeter data support the physiological concept of the surface moss forming a unit hydrologically separated from the mineral soil. Flow into the mineral soil occurs only when the organic layer is saturated (for instance during

snowmelt), while outflow (loss) through evapotranspiration by trees and shrubs continues throughout the summer. These results imply that changes in the moisture content of mineral soil over the summer may be analysed without reference to the hydrodynamics of the organic layer.

4.2.2 Summer regime of mineral soil.

The study sites are located in terrain with little local relief, except at site 1 (a mud hummock) and site 5 (in a slight depression). The surface topography suggests that there is little throughflow at the sites in summer. As a result, water movement in the active layer is predominantly vertical uptake by trees and plants and downward flow into frozen soil.

Permuzina (1978) and others (e.g. Wright 1981; Cheng 1982; Mackay 1983a) have documented downward water movement into frozen parts of the active layer and permafrost in summer (see pp. 44-48). At Mayo, downward moisture migration into frozen ground was not observed at sites 2; 4, 7 or 8. This may be due to the dry conditions at sites 2 and 4; the low unfrozen water content of the coarse-grained soil at site 7; and, at site 8, to both the coarseness of the soil and a weak temperature gradient while the seasonal frost thawed.

However, Burn and Smith (1985a) have reported neutron probe measurements of a sequential increase in the water content of soil layers at site 1 between 1 June and 30 August 1984. These measurements are reproduced in Figure 4.2. Figure 4.3 displays similar

Figure 4.2. Water content changes in successive soil layers at site 1, summer 1984. (Burn and Smith 1985a, Fig. 3).

Figure 4.3. Water content changes in the lower active layer and upper permafrost at site 3, summer 1984 and 1985.

observations made at site 3 in 1984 and 1985. In this case the data for the layer 85 - 100 cm represent downward water movement into permafrost. Downward water movement into frozen ground ahead of the thawing front was also observed at sites 5 and 6.

Figures 4.2 and 4.3 illustrate the occurrence of successive increases in moisture content with depth, as suggested by Mackay (1983a). They also indicate that the water content of certain layers increases while layers above are still frozen or themselves absorbing water. For instance, between 1 and 28 June the water content of layer 2 (65 - 80 cm) in Figure 4.2 rose by 0.18 cm^3 while the moisture content of the layer below increased by 0.13 cm^3 (Burn and Smith 1985a). While these values are close to the resolution of the neutron probe, the consistent trend in the readings indicates that they are real. Similar observations from sites 3 and 6 are reported in Table 4.4.

Temperature-induced water movement in saturated, frozen soils may result in ice segregation and frost heave (see pp. 39ff). Frost heave, therefore, is an index of the water accumulation in soil layers. Although the heave tubes installed at each site had jammed by June 1984, and in most cases did not heave during the summer, the total heave recorded by the tubes at site 1 and by the magnets reflects the water movement implied by the neutron probe measurements (Table 4.5). For instance, between 1 June and 10 July 1984 the heave tubes in the active layer at site 1 rose 7.3 mm with respect to those in permafrost. Simultaneously, the total water content increased by 6.75 mm, equi-

Table 4.4

Water content increments (cm) in various layers at sites
1, 3 and 6 in early summer.

(Data for site 1 from Burn and Smith 1985a, Table 1).

Site 1, 1984.

Depth (cm)	29/5 - 14/6	14/6 - 28/6	28/6 - 10/7
50 - 65	0.24	-	-
65 - 80	0	0.15	-
80 - 95	0	0.10	0.12

Site 3, 1984.

Depth (cm)	1/6 - 11/6	11/6 - 28/6	28/6 - 26/7
55 - 70	0.14	-	-
70 - 85	0.54	0.34	0.27
85 - 90	0.20	0.14	0.41

Site 6, 1985.

Depth (cm)	4/6 - 13/6	13/6 - 27/6	27/6 - 11/7	11/7 - 1/8
45 - 60	0.20	0.20	-	-
60 - 75	0	0.20	0.14	-
75 - 90	0	0.14	0.20	0.20

Table 4.5

Comparison of downward water movement measured by neutron probe
with summer heave data (cm).

Site	1984 ^a		1985 ^b		Magnets
	Neutron Probe	Heave Tubes	Neutron Probe	Heave Tubes	
1	0.67	0.73	Obs ^c	0.72	1.3
2	0	0	0	0	-
3	2.38	0	0.95	0	0.25
4	0	0	0	0	-
5	0.27 ^d	0	Obs	0	0.85
6	1.36	0.11	1.28	0	0.55
7	0.54	0	0	0	-
8	0	0	0	0	0

a. Cumulative water content increments in frozen soil and heave,
29 May - 30 August.

b. Cumulative water content increments in frozen soil and heave,
30 May - 1 August.

c. Increment obscured by effect of phase change on neutron
probe reading.

d. 26 July - 30 August.

valent to 7.4 mm of heave.

Summer measurements of soil displacement with the magnets may have underestimated total soil heave, since the 20 - 25 cm spacing between magnets often included both thawing (settling) and heaving sections. The closest agreement between magnets and neutron probe data was obtained at site 6, where the magnets were spaced 15 cm or less apart (see Table 4.6).

Burn and Smith (1985a) have presented determinations of K_f for the silty clay at site 1 using equation 2.11. A minimum estimate of the water flow in a soil layer is provided by the accumulation in it, since continuity dictates that:

$$\text{Inflow} - \text{Outflow} = \text{Change in storage (accumulation)}.$$

The potential gradient was computed from the temperature gradient, using a factor of $1.21 \times 10^2 \text{ m (water head) } ^\circ\text{C}^{-1}$ (p. 22). Estimates of K_f for the silty clay loam and silty clay at sites 1, 3 and 6 are presented in Table 4.7. These values may be compared with those obtained by Smith (1985a) for silty clay at Inuvik, N.W.T., which are of the same order-of-magnitude ($10^{-11} \text{ m s}^{-1}$ for -0.2 to -0.6°C ; $10^{-12} \text{ m s}^{-1}$ for -0.9 to -1.3°C).

Although water movement into permafrost was observed at site 3 from mid-July, neutron probe measurements at sites 1, 5 and 6 indicate moisture movement into permafrost only at the end of the summer. In all cases, the bulk of water movement into permafrost occurred two or three weeks before and after the start of freeze-back.

Table 4.6

Increase in water content and accompanying heave (cm),
site 6, 1985.

Water content increase, by layer.

Depth	4/6 - 13/6	13/6 - 27/6	27/6 - 11/7	11/7 - 1/8
45 - 60	0.2	0.2	-	-
60 - 75	0	0.2	0.14	-
75 - 90	0	0.14	0.2	0.2

Accompanying heave from magnets.

Depth	4/6 - 13/6	13/6 - 27/6	27/6 - 11/7	11/7 - 1/8
45 - 53	0.1	-	-	-
53 - 66	0	0.1	0.1	-
66 - 81	0	0.15	0.1	0.4

Table 4.7

Determinations of K_f from downward movement of water into
frozen ground.

(Data for site 1, 1984, from Burn and Smith 1985a, Table 2).

Date	Layer (cm)	T (°C)	dT/dz (°C m ⁻¹)	Q (cm)	K_f (m s ⁻¹)
Site 1, 1984					
29/5-14/6	50-65	-0.9	1.67	0.24	8.0×10^{-12}
14/6-28/6	65-80	-0.5	0.5	0.15	2.1×10^{-11}
14/6-28/6	80-95	-0.8	0.3	0.10	2.2×10^{-11}
28/6-10/7	80-95	-0.4	0.7	0.12	1.4×10^{-11}
Site 3, 1984.					
29/5- 4/6	85-100	-0.40	0.66	0.14	3.4×10^{-11}
4/6-11/6	70-85	-0.2	0.9	0.34	5.4×10^{-11}
11/6-21/6	70-85	-0.15	0.8	0.20	2.5×10^{-11}
21/6-28/6	70-85	-0.14	0.9	0.14	2.1×10^{-11}
1/7- 9/7	85-100	-0.3	1.0	0.14	1.5×10^{-11}
9/8-27/8	85-100	-0.25	2.0	0.34	9.0×10^{-12}
Site 3, 1985.					
13/6-20/6	70-85	-0.2	1.6	0.14	1.2×10^{-11}
4/7-11/7	70-85	-0.15	1.8	0.14	1.2×10^{-11}

Table 4.7 (cont.)

Date	Layer	T	dT/dz	Q	K_f
	(cm)	(°C)	(°C m ⁻¹)	(cm)	(m s ⁻¹)
Site 6, 1984.					
25/6-28/6	75-90	-0.5	2.93	0.14	1.5×10^{-11}
28/6- 2/7	75-90	-0.4	3.3	0.34	2.5×10^{-11}
2/7- 5/7	75-90	-0.3	3.4	0.14	1.7×10^{-11}
Site 6, 1985.					
13/6-27/6	60-75	-0.5	2.2	0.20	6.3×10^{-12}
13/6-27/6	75-90	-0.7	1.5	0.14	6.4×10^{-12}
27/6-11/7	60-75	-0.2	1.6	0.14	6.0×10^{-12}
27/6-11/7	75-90	-0.45	1.6	0.20	8.0×10^{-12}

4.2.3 Freeze-back.

Net soil moisture redistribution during freeze-back followed the well-documented pattern of desiccation in the lower active layer and enrichment closer to the ground surface (e.g. see several papers cited by Parmuzina 1978, p. 136; Mackay et al. 1979; Mackay 1980, 1983a; Cheng 1982). The total moisture redistribution at four sites over the freeze-back period is presented in Figure 4.4. These data were determined from neutron probe readings, which were corrected as described above (p. 106). As expected, greater redistribution was observed at wetter sites (e.g. site 6) than at drier ones (e.g. site 4). Most water movement during freeze-back was upward, in response to the descending frost front. However, downward water movement into permafrost was also observed at some sites, just before and during the short period of upfreezing.

In Table 4.8, the change in water content of the upper permafrost layers observed between mid-September and the completion of freeze-back is shown. Thawing of permafrost at sites 1 and 5 led to net desiccation. Deepening of the active layer at these sites may have been caused by damage to the ground surface during equipment installation.

Most water movement into permafrost was observed at the wettest site (3), with a smaller amount at sites 2 and 7. Little water movement into permafrost was observed at site 4 due to the dry conditions. At site 6, the apparent desiccation may be due to irregularities in the permafrost table leading to error in

Figure 4.4. Redistribution of moisture within the active layer
at four sites, 11 October - 20 December 1984.

Table 4.8

Net change in water content (cm^3) of permafrost from mid-September until the completion of freeze-back, 1984-85.

Site	Layer ^a	Change in θ_u^b (cm)	Completion of freeze-back
1	-5 - 10 ^c	-0.67	20/12/84
	10 - 25	0.00	
2	0 - 15	0.20	20/12/84
	15 - 30	0.14	
3	0 - 15	0.75	1/ 5/85
	15 - 30	0.00	
4	0 - 15	0.00	20/12/84
	15 - 30	0.00	
5	0 - 15	-0.82	12/ 1/85
	15 - 30	0.00	
6	0 - 15	-0.78	20/12/84
	15 - 30	0.20	
7	0 - 15	0.27	20/12/84
	15 - 30	0.00	

a. Depth (cm) below permafrost table (0 cm).

b. Increase (+) in water content of 15-cm layer.

c. 5 cm above permafrost table to 10 cm below.

determining the depth at which the neutron probe access tube encountered permafrost. The site had a surface relief of over 20 cm. It is pertinent to note that water accumulation was observed in permafrost below the desiccating layer at this site.

The total soil displacement at each site during freeze-back was similar in both years of the study. It is likely, therefore, that the frost heaving regimes at each site were similar each year. Detailed measurements of soil displacement between 9 November 1984 and 15 April 1985 were provided at sites 1, 3, 5, 6, and 8 by the magnets installed in summer 1984. As demonstrated in Table 3.8 (p. 111), soil displacements measured by the magnets and heave tubes were in close agreement.

Soil displacements in the active layer at site 1 between 9 November and 20 December, are presented in Table 4.9. Ground temperature profiles for the same period are displayed in Figure 4.5. The values in Table 4.9 are not direct measurements of segregational frost heave, since they include soil expansion due to freezing of pore water in situ. Therefore, in order to determine segregational heave, each displacement value must be corrected for the freezing of pore water, as determined using the unfrozen water content characteristic curve for the soil (see Figure 3.9, p. 79). At site 1, all displacement, except between 21 and 25 November in the 20 to 39 cm layer, may be accounted for by interstitial freezing.

However, nearly 12 cm of heave was observed between 9 November and 15 April at site 3, where the saturated soil and slow

Table 4.9

Soil displacement (cm) measured by magnets in the active layer
at site 1, 9 November - 20 December 1984.

Depth	Date			
	9/11-14/11	14/11-21/11	21/11-25/11	25/11-20/12
20 - 39	0.2	0	0.1	0
39 - 58	0.15	0.1	0	0.05
58 - 66	0	0	0	0
66 - 92	0	-0.1	0	0

Figure 4.5.. Ground temperature profiles during freeze-back
at sites 1 and 3, November 1984 - January 1985.

freezing rate produced an optimal environment for temperature-induced water migration and ice segregation. Furthermore, at this site relative displacement of several magnets was observed simultaneously, providing evidence of secondary frost heave, i.e. heave in soil on the cold side of the primary ice lens.

The magnet displacement and segregational frost heave observed at site 3 over the 1984/85 winter are presented in Table 4.10. Ground temperature profiles for the period are presented in Figure 4.5. Measurements marked with an asterisk in Table 4.10 denote secondary frost heave. These data are conservative estimates of secondary heave, since they refer to the displacement of magnets which, in some cases, were some distance from the frost front. However, since primary and secondary heave cannot be distinguished in the separation of the lowermost heaving magnet pair, they are described together as primary heave.

The data demonstrate that moisture movement continues in frozen ground in the presence of a temperature gradient. At this site saturated conditions in the unfrozen active layer provided an unrestricted water supply for ice segregation, and secondary heave was observed as clearly as in other studies (e.g. Parmuzina 1978; Smith 1985a) where the moisture supply was exhausted. Secondary frost heave (2.75 cm) accounted for at least 23 percent of the total heave at site 3 over the winter.

Measurements of frost heave at other sites are summarized in Table 4.11. At both sites 5 and 6 secondary heave was observed, with more at site 5 due to the wetter conditions. Relatively

Table 4.10

Soil displacement (cm) and frost heave (cm) measured by magnets
at site 3, 11 October 1984 - 15 April 1985.

Displacement Date	Depth (cm)			
	0 - 21	21 - 44	44 - 64	64 - 82
11/10- 9/11	2.95	0	0	0
9/11-12/11	0	0.35	0	0
12/11-14/11	0.05	0.25	0	0
14/11-21/11	0.10	0.10	0	-0.10
21/11-24/11	0	0.40	0	0
24/11-26/11	0	0.10	0	0
26/11-20/12	0	3.00	0	0
20/12-23/ 1	0	0.35	2.95	-0.20
23/ 1-29/ 1	0	0	0.25	0
29/ 1-15/4	0.05	0.10	2.30	0.45
Segregational heave				
11/10- 9/11	2.50	0	0	0
9/11-12/11	0	0.30	0	0
12/11-14/11	0	0.15	0	0
14/11-21/11	0.10*	0.05	0	0
21/11-24/11	0	0.33	0	0
24/11-26/11	0	0.10	0	0
26/11-20/12	0	2.80	0	0
20/12-23/ 1	0	0.35*	2.84	0
23/ 1-29/ 1	0	0	0.25	0
29/ 1-15/4	0	0.05*	2.00*	0.15

* denotes secondary heave.

Table 4.11

Segregational frost heave (cm) at sites 5, 6 and 8 measured with magnets, 9 November 1984 - 15 April 1985.

Site 5.

Date	Depth (cm)			
	12 - 23	23 - 33	33 - 47	47 - 60
11/10- 9/11	1.80	1.05	0	0
9/11-12/11	0	0.27	0	0
12/11-15/11	0	0.65	0	0
15/11-19/11	0	0.10	0.34	0
19/11-22/11	0.10*	0	0.20	0
22/11-26/11	0	0	0.17	0
26/11-20/12	0	0	0	0

Site 6.

	30 - 42	42 - 52	52 - 65
9/11-12/11	0.26	0	0
12/11-15/11	0	0	0
15/11-19/11	0	0.1	0
19/11-22/11	0.05*	0	0
22/11-26/11	0	0	0
26/11-20/12	0	0	0.05

Site 8.

	38 - 61	61 - 81	81 - 103
15/11-20/11	0	0.15	0
20/11-25/11	0.15*	0	0
25/11-24/12	0.05*	0.30	0.10
24/12-23/ 1	0	0	0.45
23/ 1-15/4	0	0	0.15

* denotes secondary heave.

little heave was recorded at site 8, since the soil was relatively dry. However, 0.2 cm of secondary heave was observed in the 38 to 61 cm interval between 20 November and 24 December. The neutron probe indicated slight moisture redistribution throughout the winter at this site.

Smith (1985a) has presented a method similar to that outlined above (p.136) for determining K_f using frost heave observations of field soils at temperatures well below those practicable in the laboratory. This method was used earlier to determine values for K_f from measurements of downward water movement in summer. It may also be applied to heave measurements made during freeze-back.

In Table 4.12 determinations of K_f from the Mayo heave measurements are presented for temperatures down to -1.6°C . These values are consistent with those in the literature for similar soils at the same temperatures (e.g. Yoneyama *et al.* 1983; Ohrai and Yamamoto 1985; Smith, 1985a), and also with the determinations made earlier (Table 4.7) from downward water migration into permafrost. They indicate the continuing permeability of soils at temperatures below 0°C .

4.2.4 Winter.

Theory would suggest and Parmuzina (1978) and Cheng (1982) have reported that moisture movement in the active layer continues upward along the temperature gradient throughout the winter. Parmuzina (1978) reported an increase in water content in the upper layers of the soil between December and mid-April at

Table 4.12

Calculation of K_f from frost heave measurements.

Site 1.

Date	Depth ^a	Displcmnt ^b	Heave ^c	T ^d	dT/dz ^e	K_f ^g
21/11-25/11	20 - 39	0.1	0.1	-1.6	3.8	6.0×10^{-12}

Site 3, Primary heave.

21/11-24/11	21 - 44	0.4	0.33	-0.3	1.5	6.7×10^{-11}
24/11-26/11	21 - 44	0.1	0.1	-0.2	2.0	2.4×10^{-11}
26/11-20/12	21 - 44	3.2	2.8	-0.4	2.0	5.6×10^{-11}
20/12-23/1	44 - 64	2.95	2.84	-0.5	1.3	6.4×10^{-11}

Site 3, Secondary heave.

14/11-21/11	0 - 21	0.1	0.1	-1.0	2.3	5.9×10^{-12}
20/12-23/1	21 - 44	0.35	0.3	-0.6	1.0	8.1×10^{-12}
23/1-29/1	44 - 64	0.25	0.25	-0.45	2.1	1.9×10^{-11}
29/1-15/4	44 - 64	2.3	2.00	-1.0	3.3	5.0×10^{-12}

a. Depth below surface. (cm)

b. Soil displacement. (cm)

c. Frost heave. (cm)

d. Temperature of layer. ($^{\circ}\text{C}$)e. Temperature gradient through layer. ($^{\circ}\text{C m}^{-1}$)g. Hydraulic conductivity. (m s^{-1})

Table 4.12 (cont.)

Site 5.

Date	Depth ^a	Displcmnt ^b	Heave ^c	T ^d	dT/dz ^e	K _f ^g
1/11-9/11	12 - 23	1.9	1.8	-0.2	2.0	9.0×10^{-11}
9/11-12/11	23 - 33	0.3	0.27	-0.1	1.1	1.1×10^{-11}
12/11-15/11	23 - 33	0.7	0.65	-0.3	3.3	6.2×10^{-11}
15/11-19/11	33 - 47	0.4	0.34	-0.1	1.6	5.1×10^{-11}
19/11-22/11	33 - 47	0.2	0.2	-0.1	1.3	4.9×10^{-11}
24/11-26/11	33 - 47	0.2	0.17	-0.15	1.9	4.3×10^{-11}

Site 6.

9/11-12/11	30 - 42	0.4	0.26	-0.6	3.8	2.2×10^{-11}
15/11-19/11	42 - 52	0.15	0.10	-1.0	3.4	7.0×10^{-12}
19/11-22/11	30 - 42	0.1	0.05	-1.3	3.3	4.8×10^{-12}
26/11-20/12	52 - 64	0.15	0.05	-1.4	6.0	3.3×10^{-13}

Site 8.

15/11-20/11	61 - 81	0.15	0.15	-0.1	1.6	1.8×10^{-11}
20/11-25/11	38 - 61	0.15	0.15	-0.4	1.3	2.2×10^{-11}
25/11-24/12	38 - 61	0.05	0.05	-1.1	2.0	8.3×10^{-13}
25/11-24/12	61 - 81	0.45	0.35	-0.3	2.5	4.6×10^{-12}
24/12-23/1	81 - 103	0.55	0.45	-0.1	2.0	7.2×10^{-12}

a. Depth below surface. (cm)

b. Soil displacement. (cm)

c. Frost heave. (cm)

d. Temperature of layer. ($^{\circ}\text{C}$)e. Temperature gradient through layer. ($^{\circ}\text{C m}^{-1}$)g. Hydraulic conductivity. (m s^{-1})

three sites near Ust' Yeniseyek, U.S.S.R. (see p. 50).

In a similar manner, neutron probe measurements at Mayo indicate a small increase (0.2 cm) in the moisture content of the upper 15 cm of mineral soil at sites 1, 4, 6 and 7 over the winter (20 December to 15 April). A larger increment was recorded at sites 3 (1.7 cm) and 8 (0.27 cm), where "freeze-back" continued until late-March. Measurements of soil vapour outflow at sites 2 and 8 also indicate that water movement continued over the winter (cf. pp. 52-53).

The vapour outflow measurements were made in vapour traps constructed from tin cans covered by a Gortex membrane, permeable to water vapour but not to liquid. Four traps were placed on the ground in September 1984 and removed at intervals over the winter. The cans were enveloped by the snowpack after mid-November, and interfered little with the natural hydrological system. Measurements of vapour flow out of the soil between 3 October 1984 and 25 April 1985 are presented in Table 4.13.

Vapour entrapment continued throughout the winter at site 8, but terminated at site 2 by 1 April, when the temperature gradient was reduced. The total water trapped at site 8 (4.07 mm) was nearly double that at site 2 (2.19 mm). This may be due to the continuing availability of water to the seasonally frozen layer at site 8, and to surface effects induced by snowmelt at site 8 by 15 April. The neutron probe did not detect any snowmelt infiltration at site 2 until 29 April.

Although the water contributing to outflow and upper layer

Table 4.13

Vapour outflow measurements, sites 2 and 8,
October 1984 - April 1985.

Period	Days	Outflow (mm)	Rate (mm/day)
Site 2			
3 Oct. - 21 Nov.	49	0.92	0.019
21 Nov. - 23 Jan.	63	0.75	0.012
23 Jan. - 30 Mar.	66	0.48	0.007
30 Mar. - 25 Apr.	25	0.00	0.000
Site 8			
3 Oct. - 21 Nov.	49	0.95	0.019
21 Nov. - 23 Jan.	63	1.21	0.019
23 Jan. - 30 Mar.	66	1.11	0.017
30 Mar. - 25 Apr.	25	0.80	0.032

accumulation would be detectable by the neutron probe if it originated entirely from any one 15 cm layer, such was only observed at sites 3 and 8. Therefore, it is likely that this water comprises contributions from all parts of the soil profile, including permafrost (Parmuzina 1978; Cheng 1982).

As discussed earlier (pp. 52-53), Smith and Burn (in press) were unable to reconcile these measurements with the theory of vapour diffusion in soils. Instead they proposed that the flow through the soil was in the liquid phase, driven by the temperature gradient. They presented calculations for K_f of the soil, based on the outflow and the mean temperature gradient during each interval (see Table 4.14). The values obtained, on the order of $10^{-13} \text{ m s}^{-1}$, are consistent with others reported in the literature for fine-grained soils at these temperatures (e.g. Ohrai and Yamamoto 1985, Smith 1985a). While lower soil temperatures were recorded at site 2, values determined for K_f are similar to those from site 8. This may be explained by the difference in soil type, which leads to a higher θ_u at site 2 (cf. Figures 3.9, 3.12).

4.2.5 Snowmelt.

When snowmelt begins, percolation and refreezing of meltwater within the snowpack bring the snow rapidly to an isothermal state at 0°C (Marsh and Woo 1984). Conditions favouring temperature-induced infiltration into the soil are established, with cold soil beneath a "warm", wet snowpack (Kane and Stein 1983a; Burn and

Table 4.14

Values of apparent hydraulic conductivity determined from
vapour outflow measurements.

(Smith and Burn, in press, Table 4).

Period	Q (g cm ⁻² day ⁻¹)	dT/dz (°C cm ⁻¹)	T ₁₀ ^a (°C)	K _f (m s ⁻¹)
Site 2				
3 Oct. - 21 Nov.	.0019	0.0320	-2.5	5.7 x 10 ⁻¹³
21 Nov. - 23 Jan.	.0012	0.0290	-5.5	4.0 x 10 ⁻¹³
23 Jan. - 30 Mar.	.0007	0.0355	-6.7	1.9 x 10 ⁻¹³
30 Mar. - 25 Apr.	.0	0.0	-	-
Site 8				
3 Oct. - 21 Nov.	.0019	0.0615	-0.3	3.0 x 10 ⁻¹³
21 Nov. - 23 Jan.	.0019	0.0680	-2.3	2.7 x 10 ⁻¹³
23 Jan. - 30 Mar.	.0017	0.0500	-2.5	3.3 x 10 ⁻¹³
30 Mar. - 25 Apr.	.0032	0.0055	-1.0	5.6 x 10 ⁻¹²

a. Mean soil temperature at a depth of 10 cm.

Smith 1985b). As mentioned above (p. 44), Kane and Stein (1983b) indicate that most groundwater recharge in central Alaska occurs during snowmelt.

Snowmelt infiltration was monitored closely at the study sites between mid-April and 3 May 1985. Net increases in soil water content attributable to ~~snowmelt~~ infiltration were also measured between 4 April and 31 May 1984, and 3 May and 31 May 1985.

The snowmelt infiltration at each study site in April and early May 1985 is presented in Figure 4.6. Infiltration was observed at all sites except site 4; but first at site 8 (between 10 and 15 April) and last at site 2 (between 27 and 29 April). Snowmelt infiltrated into the upper 45 cm of the soil by 2 May, and at the end of May each year an increase in soil water content associated with snowmelt was noticed between 60 and 75 cm at site 2. The progressive moisture content increases at sites 1 and 6 displayed in Figure 4.7 clearly indicate the ability of frozen soil to conduct water.

The soil recharge recorded at most sites was on the order of a few centimetres. Little infiltration was observed at site 2 by 2 May due to the late start of snowmelt; at site 3 due to saturated soil conditions coupled with a weak temperature gradient; and at site 4 due to the shallow snowpack. Overall, snowmelt was the major recharge period of the year (cf. Kane and Stein 1983b).

Conventionally, snowmelt infiltration has been attributed to gravity (e.g. Kane and Stein 1983a). However, Burn and Smith

Figure 4.6. Snowmelt infiltration and accompanying ground warming at the study sites, 24 April - 2 May 1985.

Figure 4.3. Pattern of water content increases in the surface
0-15 layers of sites 1 and 6, 24 April - 2 May 1985.

(1985b) have suggested that temperature-induced potential gradients are important for snowmelt uptake by frozen soil. The data from site 3 support this suggestion, since the soil at this site was close to saturation. Nevertheless, infiltration into the organic layer is driven by gravity, as the appropriate potentials for temperature-induced flow are not generated in such a medium. Much of the infiltration at sites 1, 5, 6 and 7 was recorded in the organic layer, where the high porosity accommodated the water.

In 1985, snowmelt infiltration was generally accompanied by soil heave (Table 4.15). The limited heave compared to infiltration indicates that at most sites recharge occupied soil pore space rather than ice lenses. However, at site 3 about one quarter of the infiltration resulted in heave, and at site 8 the infiltration first replenished the soil moisture deficit and then contributed to heave.

Woo and Heron (1981), Wright (1981) and Marsh and Woo (1984) have suggested that the latent heat released by snowmelt as it refreezes in the soil contributes significantly to soil warming (see temperature profiles in Figure 4.6). This occurs principally below about -1.2°C , since pore ice melting is initiated at temperatures above this in these soils (see Figure 3.9). At the warmer temperatures most of the latent heat released by snowmelt refreezing is used to thaw pore ice rather than warm the soil. It is important to note that in very dry soil some of the infiltration may not freeze, but instead fulfill the unfrozen water

Table 4.15

Snowmelt infiltration and accompanying soil heave,
mid April - 2 May 1985.

Site	Snowmelt infiltration (cm)	Heave from heave tubes (cm)	Heave from magnets (cm)
1	2.33	0	0.2
2	0.27	0	-
3	0.88	0.2	0.2
4	0	0	-
5	2.72	0	0.05
6	2.94	0	0.5
7	1.98	0	-
8	7.30	0.42	0.75

content requirement of the soil,

4.3 Annual water balance.

The purpose of this thesis is to document the aspects of the hydrologic regime that were identified by Cheng (1982) as contributing to the accumulation of ice in the upper layer of permafrost. The discussion of seasonal moisture fluxes presented above deals with some of these aspects. Cheng (1982) proposed that aggradational ice was the product of the integration of these flows over the year.

In this section, therefore, the annual water balance at the study sites is presented, as determined with the neutron probe, heave tubes and soil analogue tubes. The data obtained at sites 1 and 5 are not considered in detail, since damage to the surface during equipment installation induced thawing of permafrost.

4.3.1 Results from the neutron probe.

Some of the results in section 4.2 are complicated by the effects of freezing and thawing on the neutron probe. These complications are avoided if changes of water content in the soil profile are determined between dates when ground temperatures are similar. Since such conditions occur at yearly intervals, the annual water balance may be determined within the rated accuracy of the neutron probe.

The data in Table 4.16a indicate the water balance at sites 1 to 7 over periods of a year between 8 August 1983 and 8 October

Table 4.16a

Annual water balance (cm^3) at sites 1 to 7,

8 August 1983 - 9 October 1985.

	8 Aug 83- 9 Aug 84.	7 June 84- 7 June 85.	25 July 84- 26 July 85.	9 Oct 84- 8 Oct 85.
Site 1				
Active layer	0	+2.6	+0.3	+0.2
Permafrost	-0.3	-1.1	0	-0.2
Total	-0.3	+1.6	+0.3	0
Site 2				
Active layer	+3.3	+0.7	+3.9	-0.3
Permafrost	+0.3	0	-1.2	+0.8
Total	+3.6	+0.7	+2.7	+0.5
Site 3				
Active layer	+0.3	-0.1	+0.8	+0.3
Permafrost	+1.4	+0.7	-0.7	0
Total	+1.7	+0.6	+2.7	+0.3
Site 4				
Active layer	-4.6	-4.3	-	-
Permafrost	+0.2	0	-	-
Total	-4.4	-4.3	-	-
Site 5				
Active layer	-	-	+1.6	-1.8
Permafrost	-	-	-1.4	-0.5
Total	-	-	+0.2	-2.3

Table 4.16a (cont.)

Annual water balance (cm^3) at sites 1 to 7,

8 August 1983 - 9 October 1985.

	8 Aug 83- 9 Aug 84.	7 June 84- 7 June 85.	25 July 84- 26 July 85.	9 Oct 84- 8 Oct 85.
Site 6				
Active layer	-	-5.7	-2.6	+3.9
Permafrost	-	-1.5	+0.2	+0.3
Total	-	-7.2	-2.4	+4.2
Site 7				
Active layer	-5.4	-4.8	-1.6	-1.0
Permafrost	+0.5	0	0	-0.2
Total	-4.9	-4.8	-1.6	-1.2

Table 4.16b

Water balance (cm^3) of sites with two years' record,

8 August 1983 - 25 July 1985.

	Site 1	Site 2	Site 3	Site 7
Active layer	+0.3	+7.2	+1.1	-7.0
Permafrost	-0.3	-0.9	+0.7	+0.5
Total	0	+6.3	+1.8	-6.5

1985. The data refer to the water balance of mineral soil, since the surface organic layer is independent of the mineral soil-water system (p. 125). The profile has been divided into the active layer and permafrost, in order to isolate the component entering or leaving permafrost, which is of ultimate interest to this study.

The data presented in Table 4.16a were calculated from the mean readings over a fortnight at each end of the interval. This is to accommodate the variation in active-layer development on a daily basis from summer to summer. For instance, readings taken on 1, 8 and 16 August 1983 were used to calculate the moisture content representing the beginning of the period 8 August 1983 to 9 August 1984.

Predictably, Table 4.16a indicates that the hydrological regime of the active layer is more dynamic and more variable than that of permafrost. However, the data indicate that the water content of permafrost generally increased over the period. Changes in the moisture content of permafrost at most sites were close to the resolution of the neutron probe (± 0.2 cm over a 15 cm interval), which may explain some of the scatter in the observations. Nevertheless, while the largest changes in permafrost water content per se were observed at the damaged sites (in the uppermost 15 cm of permafrost in each case), the most consistent increase was recorded at site 3, where conditions were optimal for temperature-induced water movement. Over the two-year period beginning in August 1983 (see Table 4.16b), the mean rate

of moisture incorporation into permafrost at this site was 3.5 mm year⁻¹.

The active layer did not behave in a similar manner at all sites. Net desiccation of the active layer was observed at site 4, due to low snow accumulation (a temporary effect induced by the thaw-slump headwall exposing the site), and at site 7. At the other sites there was little change, except for site 2, where a large increase in moisture content was recorded over two years beginning in August 1983.

The data indicate that the water balance of permafrost terrain varies considerably from year to year. Changes in moisture content of permafrost always occurred within 30 cm of the bottom of the active layer. Below this depth, any changes were smaller than the resolution of the neutron probe. The maximum annual increase in the water content of permafrost was two orders-of-magnitude less than mean annual total precipitation in the Mayo area (306 mm).

4.3.2 Net heave of uppermost permafrost layer.

Annual displacements of the lowermost two heave tubes at the sites with permafrost between 1 August 1983 and 8 October 1985 are shown in Table 4.17. During this period, consistent heave in permafrost was only recorded at sites 1 and 3. At site 2 all of the heave occurred during freeze-back in fall 1983, and may be an

Table 4.17

Heave (+) and subsidence (-) (cm)^a of heave tubes in permafrost,
sites 1 - 7, August 1983 - October 1985.

Site & Depth (cm)	1 Aug. 83- 1 Aug. 84.	1 Aug. 84- 1 Aug. 85.	1 Aug. 83- 1 Aug. 85.	13 Sept. 83- 10 Oct. 85.
Site 1				
100 - 125	+0.10	+0.04	+0.14	+0.15
125 - 150	+0.03	0	+0.03	+0.07
Site 2				
100 - 125	0	0	0	0
125 - 150	+0.60	0	+0.60	+0.60
Site 3				
100 - 125	0	+4.20	+4.20	0
125 - 150	-0.19	+0.70	+0.51	+0.69
Site 4				
100 - 125	0	0 ^b	-	-
125 - 150	+0.04	0 ^b	-	-
Site 5				
100 - 125	0	0	0	0
125 - 150	0	0	0	0
Site 6				
100 - 125	0	+2.89	+2.89	+2.92
125 - 150	0	0	0	0
Site 7				
100 - 125	0	0	0	0
125 - 150	0	0	0	0

a. Resolution is ± 0.02 cm.

b. Heave between 1 Aug. 84 and 27 June 85.

artifact of the installation, since no accompanying or further displacement was measured in any part of the profile at this site throughout the study. The large heave measured in the second year by the upper heave tube in permafrost at sites 3 and 6 is probably due to jacking of the tubes, as described above (p. 109). Overall, the heave measurements compare reasonably well with the neutron probe determinations discussed above, especially at site 3, where the 6.9 mm heave observed between 13 Sept 1983 and 10 Oct 1985 is in close agreement with the average rate of 3.5 mm year^{-1} measured with the neutron probe.

Due to the inadequacies of the initial measuring device, the soil displacement magnets did not provide any reliable indication of heave in permafrost over a year. It is hoped this problem has now been overcome, and that they present considerable potential for continuing observations.

4.3.3 Soil analogue tubes.

While soil conditions at several of the sites in this study were relatively dry, e.g. sites 2, 4 and 7, the silt loam in the closed soil analogue tubes represented optimal conditions for temperature-induced water movement at all sites. The tubes were originally filled with super-saturated soil, which settled over the study period. Water was drained from the top of each tube in October 1984 and 1985 after settlement over the summer. The soil remained saturated throughout the study.

In October 1984, ice lenses were observed in most tubes just

below the base of the active layer (e.g. Figure 4.8). These ice lenses grew over the following year, and by October 1985 they were up to 1.5 cm thick. Ice lens growth was accompanied by an increase in the water content of the appropriate 10 cm sections of tubes, as shown by samples collected at intervals between 2 August 1984 and 10 October 1985.

Changes in the water content of the section of the tube containing the upper 10 cm of permafrost at 5 sites are displayed in Figure 4.9. The data exhibit some scatter due to variations in the original filling of the tubes. However, an increase in the moisture content of this zone is observed at each site over the period concerned. The trend is accentuated when the mean moisture content of the uppermost 20 cm of permafrost is examined (see Figure 4.10).

The net increases in water content of the uppermost permafrost zone in the tubes over the periods 2 August 1984 to 24 July 1985 and 5 October 1984 to 10 October 1985 are shown in Table 4.18. These data may be considered as estimates of the maximum possible net water migration into permafrost under the thermal regimes found in the study area, since there was no moisture stress in the tubes and the soil was particularly frost-susceptible.

The dry density of the silt loam in the tubes was determined in the laboratory to be 1.2 g cm^{-3} , close to the dry density of the natural soil. Combining this value with the mean gravimetric water content increases reported in Table 4.18, upper limits of

Figure 4.8. Ice lens in the upper 10 cm of permafrost in
a soil analogue tube at site 2. 27 July 1985.
(The scale is in intervals of 5 cm.)

Figure 4.9. Increase in water content of the upper 10 cm of permafrost in the soil analogue tubes at the study sites between 2 August 1984 and 10 October 1985.

Figure 4.10. Increase in water content of the upper 20 cm of permafrost in the soil analogue tubes at the study sites between 2 August 1984 and 10 October 1985.

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MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS
STANDARD REFERENCE MATERIAL 1010a
(ANSI and ISO TEST CHART No. 2)

Table 4.18

Increase in gravimetric water content (g g^{-1}) of the uppermost
permafrost layer in the soil analogue tubes,
August 1984 - October 1985.

Site	Depth (cm)	2 August 84- 24 July 85.)	5 October 84- 10 October 85.
1	110 - 120	0.08	0.12
1	110 - 130	0.06	0.09
2	90 - 100	0.17	0.07
2	90 - 110	0.11	0.06
3	80 - 90	0.06	0.03
3	80 - 100	0.05	0.02
4	90 - 100	0.01 ^a	-
4	90 - 110	0.05 ^a	-
7	80 - 90	0	0.01
7	80 - 100	0	0.01
Mean	Upper 10 cm of permafrost	0.06	0.06
Mean	Upper 20 cm of permafrost	0.05	0.04

a. Increase between 2 August 1984 and 18 June 1985.

$1.2 \text{ cm}^3 \text{ year}^{-1}$ and $0.96 \text{ cm}^3 \text{ year}^{-1}$ are estimated for the mean rate of water incorporation into the uppermost 20 cm of permafrost. A slightly lower value ($0.7 \text{ cm}^3 \text{ year}^{-1}$) may be calculated for annual water accumulation in the upper 10 cm of permafrost. These rates may be compared with the moisture movement into permafrost observed at site 3 ($0.35 \text{ cm}^3 \text{ year}^{-1}$, Table 4.16b).

4.4 The Cheng hypothesis.

The results presented in this chapter support the Cheng hypothesis for the origin of aggradational ice in permafrost. Downward water migration into permafrost was observed at several sites in summer, while relatively little desiccation occurred in winter.

A net increase in the water content of permafrost over the study period was measured by the neutron probe at sites 3 and 7, and over the second year of the study at site 2. Decreases in the water content of the upper permafrost zone at sites 1 and 5 may be associated with the disturbance to surface conditions wrought during site installation. The largest annual increase was observed at site 3 (1.4 cm^3) between August 1983 and 1984. At this site, conditions were the most suitable for temperature-induced water movement, and, hence, for the processes considered by the hypothesis.

Measurements of soil heave made with the telescoping heave tubes also support the hypothesis: both heave due to downward water migration in summer, and an annual heave in the uppermost

layers of permafrost was observed. No subsidence was observed in the permafrost layers over the study period.

A vivid demonstration of the hypothesis was presented by the development of ice lenses in the uppermost permafrost layers of the soil analogue tubes. These observations clearly illustrate progressive water movement into permafrost without ground cracking. Determinations of a maximum rate of water migration into permafrost on the order of $1 \text{ cm}^3 \text{ year}^{-1}$ were provided by these tube experiments, in agreement with the increase observed at site 3 between August 1983 and 1985. The data suggest that the processes considered by the hypothesis result in a net incorporation of water into permafrost when it is in equilibrium or aggrading. Because of the small quantities of moisture involved, a small deepening of the active layer may lead to the loss of several years' moisture accumulation from permafrost in one season. The extent of this was observed at sites 1 and 5.

CHAPTER 5

ADDITIONAL FIELD OBSERVATIONS

Two investigations of permafrost hydrogeology were conducted in the study area to examine the Cheng hypothesis over longer time periods than considered in Chapter 4. First, recent water migration into permafrost was assessed by examining the infiltration of tritiated water ($^3\text{H}_2\text{O}$), naturally precipitated after the atmospheric nuclear tests of the 1950s and early 1960s. Second, the ground ice stratigraphy exposed in the thaw slumps was examined for evidence regarding the hypothesis over geological time.

5.1 Migration of tritium into permafrost.

During the atmospheric nuclear tests in the 1950s and early 1960s, tritium (^3H) concentrations in precipitation rose to a peak value several orders-of-magnitude higher than pre-test or present values. In 1963 the mean tritium concentration in precipitation at Whitehorse was over 4100 T.U. (1 T.U. = 1 ^3H in 10^{18} H atoms; R.M. Brown 1970), but is now much less. Precipitation collected for the author on a monthly basis at Whitehorse and Mayo Airports between October 1985 and March 1986 had a mean tritium concentration of less than 30 T.U.

Although tritium has a half-life of 12.26 years (Kell 1972), the exceptional tritium concentration in thirty year old ground-

water can still be detected (Michel et al. 1984). Indeed, tritium is one of the few viable tracers for studying water movement in permafrost. Whereas dissolved solutes alter the behaviour of water below 0°C , the isotopes of oxygen and hydrogen (^{16}O , ^{18}O , ^1H , ^2H , ^3H) are directly incorporated into the water molecule. However, both fractionation of each isotope during freezing and diffusion of the isotopes in bulk water and ice complicate determinations of water movement.

Fractionation is the preferential incorporation or rejection of an isotope as a medium changes phase (Suzuoki and Kimura 1973). Heavy isotopes, such as tritium, are preferentially concentrated in the lower energy ice phase as water freezes. Hence, the concentration of these isotopes in precipitation declines as distance from the ocean increases or temperature decreases. Under equilibrium conditions, i.e. when a phase change occurs relatively slowly, the amount of fractionation depends partly on the difference in atomic weight between the isotopes of an element (Michel 1986). Tritium is fractionated more than deuterium (^2H), for example. Michel (1982) frequently observed fractionation during laboratory tests investigating the slow freezing of fine-grained soil-water systems. For the purposes of this discussion, it is important to note that tritium is preferentially incorporated in ice during soil freezing, and, therefore, θ_u is depleted in this isotope.

Diffusion is the principal process maintaining an equilibrium chemical concentration in homogeneous solids, and occurs at a rate

directly proportional to the concentration gradient in the medium. The coefficient of proportionality is the diffusion coefficient, D . Convection may increase the apparent rate of diffusion in liquids and gases, and hence D increases for these phases. The diffusion coefficient of tritium in water near 0°C is $10^{-5} \text{ cm}^2 \text{ s}^{-1}$ (Wang et al. 1953), and in ice $10^{-11} \text{ cm}^2 \text{ s}^{-1}$ (Jaccard 1972).

Both Michel and Fritz (1978; 1982) and Chizhov et al. (1983) have examined the tritium content of natural permafrost. Chizhov et al. (1983, p. 149) reported that "high tritium concentrations are found...0.2 to 0.3 m below the upper boundary of permafrost", and Michel and Fritz (1978) detected tritiated water up to 0.9 m below the permafrost table at a site near Norman Wells, N.W.T. However, the latter authors did not find enriched tritium below the active layer on Richards Island, N.W.T. (Michel and Fritz 1982).

Chizhov et al. (1983) suggested that the tritium moved into permafrost by mixing at the base of a deepened active layer during warm summers and by diffusion. In order to support this interpretation, they calculated the observed "diffusion" using a coefficient of $6 \times 10^{-4} \text{ cm}^2 \text{ s}^{-1}$, which is over six orders-of-magnitude higher than that of tritium ($10^{-11} \text{ cm}^2 \text{ s}^{-1}$) or deuterium ($10^{-10} \text{ cm}^2 \text{ s}^{-1}$) in ice, and almost two orders-of-magnitude higher than D for these hydrogen species in liquid water. Wang et al. (1953) reported that the diffusion coefficients of deuterium and tritium are almost equal in bulk water; and Oliphant and Tice

(1984) have determined D for deuterium in frozen clay as $10^{-10} \text{ cm}^2 \text{ s}^{-1}$. There should not be a large difference between the diffusion coefficients of these isotopes in frozen soil. Further, Chizhov *et al.* (1983) used an annual diffusion period beginning well before much active-layer development occurred.

In this context, it is interesting to note that Michel and Fritz (1978, p. 210) observed "permafrost containing tritium (at) a greater depth in fine-grained sediments than in coarser-grained aggregates." This observation does not support mixing and diffusion in thawed soil as the process behind the tritium infiltration, since unfrozen coarse aggregate is more permeable than fine-grained sediment (Freeze and Cherry 1979). Together, Chizhov *et al.*'s (1983) calculation and Michel and Fritz's (1978) observation suggest that an alternative to diffusion is required to explain the observed infiltration of tritium into permafrost.

The next sections of this chapter contain a description of the tritium content of ground ice in the study area and experiments conducted to determine the mechanism of downward tritium movement into permafrost.

5.1.1 Tritium concentrations in permafrost, Mayo study area.

Pore water extracted from drill core samples taken at various sites in the study area was analysed for tritium concentration at the University of Waterloo. The samples were cleaned, double-bagged and weighed in the field before being sent to the laboratory for pore-water extraction. The samples were reweighed

before squeezing to determine whether evaporation had occurred in transit, but none was detected. Tritium concentrations were determined with a resolution of ± 8 T.U. Results from profiles taken near three of the experiment sites are presented in Table 5.1.

Additionally, ground ice was collected from the thaw slump headwall on 14 October 1984, about 80 m west of site 4. The ice was sampled at intervals chosen with respect to the cryostratigraphy. The surface 2 cm of the exposure was removed before samples were collected, and the samples were brought back to the laboratory frozen. A stratigraphic column of the section with the appropriate tritium determinations is presented in Figure 5.1.

The results presented in Table 5.1 indicate tritium infiltration into the upper 0.1 to 0.5 m of permafrost near sites 1, 2 and 4. (Michel and Fritz (1982) considered concentrations of about 30 T.U. or less to be physically insignificant or "background".) These results are similar to profiles presented by Chizhov *et al.* (1983) and Michel and Fritz (1978). Infiltration into permafrost was only observed at sites 1, 2 and 4, but since site 7 is of a coarse material (see Appendix 1), the results are consistent with Michel and Fritz's (1978) report of deepest infiltration in fine-grained soil. The deep tritium infiltration near sites 1 and 4 may be associated with wet soil conditions.

Figure 5.1 indicates the tritium concentration associated with different cryotextural units in the thaw-slump headwall (see Chapter 3, p. 70). Ground ice over 60 cm below the base of the

Table 5.1

Tritium concentrations (T.U.) between 50 and 150 cm depth
at 4 sites, Mayo study area.

Site 1			Site 2			Site 7			Site 4 ^a		
Depth			T.U.			T.U.			Depth		
50 - 60			-			-			240		
60 - 70			96			158			170		
70 - 80			121			137			20		
80 - 90			127			136			12		
90 - 100			-			103			18		
100 - 110			239			34			6		
110 - 120			313			27			16		
120 - 130			116			30			15		
130 - 140			20			24			5		
140 - 150			70			20			15		
									108		
									277		
									356		
									148		
									13		
									31		
									73		
									15		

Depth of permafrost table (cm)^b:

100 93 72 86

Dates of drilling: Site 1 21 June 1983

 Site 2 8 July 1984

 Site 4 9 July 1983

 Site 7 10 July 1984

a. Unpublished data from F.A. Michel, personal communication, 1985.

b. Depth of permafrost table determined from the first ice lens more than 0.3 cm thick encountered in core.

Figure 5.1. Tritium content of various cryotextural units,
Mayo thaw-slump exposure.

active layer did not contain significant quantities of tritium. However, elevated tritium levels extended into the ice-rich zone at the base of the active layer, which contained a higher concentration than soil water in the active layer. This concentration inversion is also a feature of the record from near sites 1 and 4, and provides further evidence to refute the diffusion hypothesis for tritium dispersal in permafrost.

5.1.2 Tritium movement into permafrost on an annual basis.

The low concentrations of tritium found in permafrost near Mayo may be taken to represent tritiated water moving into permafrost over the past 30 years. However, it is important to resolve whether the samples collected reflect migration of individual tritiated molecules into permafrost by diffusion, or downward bulk water movement from the active layer over the period.

In order to address this problem, two 4 m² plots near sites 2 and 7 were irrigated in mid-September 1983 with 5 litres of tritiated water (1.7×10^5 T.U.). The plots were core drilled in July 1984 and 1985 to determine the distribution of tritium within the profile. The results were compared with reference samples taken 10 m away from the plots in untreated terrain. At site 2 the reference hole was drilled upslope of the plot, and at site 7 the terrain was flat. The results are presented in Table 5.2.

An increase in the tritium concentration of permafrost after irrigation is apparent at both sites in the 1984 data. The higher

Table 5.2

Tritium concentrations (T.U.) in the lower active layer
and uppermost permafrost, reference and irrigated plots,
Mayo study area.

Depth	Site 2			Site 7		
	Reference	1984	1985	Reference	1984	1985
50 - 60				240	990	454
60 - 70	158	336	671	170	234	333
70 - 80	137	252	224	20	250	116
80 - 90	136	164	62	12	38	-
90 - 100	103	90	31	18	47	-
100 - 110	34	70	26	6	6	-
110 - 120	27	80	45	16	15	-
120 - 130	30	50	25	16	15	42
130 - 140	24	85	42	5	32	-
140 - 150	20	102	48	15	27	61

Date: 8 July 8 July 17 July 13 July 13 July 28 June

Permafrost table at 93 cm (site 2) and 72 cm (site 7).

concentration was detected throughout the uppermost 60 cm of permafrost at site 2 and 30 cm at site 7, with deeper infiltration in the fine-grained soil. However, the 1985 data do not replicate this trend and are somewhat irregular. At both irrigated plots tritium concentrations in the 1985 samples were generally lower in the permafrost, though not in the active layer. This decrease cannot be accounted for entirely by radioactive decay, since the half-life of tritium is over an order-of-magnitude longer than the sampling interval. It may be due to variations in areal concentration of tritium during initial treatment, as the ground was sprinkled rather than flooded.

In order to assess the diffusion hypothesis for these data, a diffusion coefficient was calculated (in order-of-magnitude terms) for the movement of tritium in the surface layers of permafrost, and compared to published values of this coefficient for ice, water and soil. The tritium concentration gradient was calculated from the mean concentration in the lowermost 20 cm of the active layer in each 1984 irrigated-plot core and from the concentration between 10 and 30 cm below the permafrost table in the reference core. The total flow was calculated from the difference in tritium content 10 cm or more below the permafrost table in the 1984 irrigated-plot and reference cores.

The diffusion of tritium in soil may be estimated from the diffusion equation (cf. equation 2.8):

$$Q_T = -D(\tau) \cdot \frac{d\tau}{dz} \quad (5.1),$$

where Q_T = tritium flux ($\text{mol cm}^{-2} \text{s}^{-1}$),

$D(\tau)$ = diffusion coefficient of tritium in permafrost
($\text{cm}^2 \text{s}^{-1}$),

τ = tritium concentration (mol cm^{-3}).

Both Q_T and τ are directly proportional to $\tau_a \theta_{va}$, or $\tau_p \theta_{vp}$, where τ_a , τ_p are the observed concentrations (T.U.) in the active layer and permafrost, and θ_{va} and θ_{vp} the volumetric water contents of the same layers. Q_T is obtained by summing $\tau_p \theta_{vp}$ over all intervals deeper than 10 cm in permafrost. In finite form we have:

$$\sum (\tau_p \theta_{vp} \cdot l) = -D(\tau) \cdot \frac{(\tau_a \theta_{va} - \tau_p \theta_{vp})}{\Delta z} \cdot t \quad (5.2),$$

where l = thickness of each interval.

Further, since θ_{vp} and θ_{va} are similar in order-of-magnitude terms, for our purpose we obtain:

$$\sum (\tau_p \cdot l) = -D(\tau) \cdot \frac{(\tau_a - \tau_p)}{\Delta z} \cdot t \quad (5.3).$$

(As $\theta_{va} \approx \theta_{vp}$, this simplification reduces the estimate of $D(\tau)$.)

Calculations of $D(\tau)$ for both sites are presented in Table 5.3, where laboratory determinations of tritium concentration have been corrected for radioactive decay over the period concerned. The estimates for $D(\tau)$ of $1.6 \times 10^{-5} \text{ cm}^2 \text{s}^{-1}$ (site 2) and $1.7 \times 10^{-6} \text{ cm}^2 \text{s}^{-1}$ (site 7) are over four orders-of-magnitude higher than the diffusion coefficients of tritium in ice or deuterium in frozen clay cited above. Similar calculations, using the 1985 irrigated plot data provided estimates for $D(\tau)$ of $2.5 \times 10^{-6} \text{ cm}^2 \text{s}^{-1}$ (site 2) and $1 \times 10^{-7} \text{ cm}^2 \text{s}^{-1}$ (site 7).

Table 5.3

Calculation of apparent diffusion coefficients
for tritium in permafrost.

	$\Sigma(1. r_p)^a$	r_a^b	r_p^c	z	$d r / d z$	t^d	$D(\tau)$
	TU.cm	TU	TU	cm	TU cm ⁻¹	s	cm ² s ⁻¹
Site 2.							
Determined ^e	2520	208	30	30		2.6×10^7	
Corrected ^f	2670	220	32				
Result					6.3		1.64×10^{-5}
Site 7.							
Determined	940	612	15	30		2.6×10^7	
Corrected	990	648	16				
Result					21		1.7×10^{-6}

- a. Obtained from difference between concentration in 1984 reference and irrigated-plot samples, Table 5.2.
- b. Mean concentration in lowest 20 cm of the active layer.
- c. Mean concentration 10 - 30 cm below the permafrost table, reference core.
- d. Mid-September 1983 to mid-July 1984.
- e. Laboratory determinations of tritium concentration in pore water, 12 March 1985.
- f. Concentration corrected by $e^{-0.0565t}$,
where -0.0565 = decay constant for tritium
 t = time in years.
A correction of 8 months was applied to r_p in reference core and 1 year to other concentrations.

These results indicate that the diffusion model does not explain the observed migration of tritium molecules into permafrost. Instead, it is suggested that the downward movement of tritium over the past 30 years reported here, by Chizhov *et al.* (1983) and by Michel and Fritz (1978) is due to temperature-induced water migration. In this case, the relatively deep infiltration into fine-grained sediments observed both at Mayo and by Michel and Fritz (1978), is consistent with the higher K_f of these materials compared to coarse-grained soil.

5.2 Cryostratigraphic evidence of water movement into permafrost.

The ground ice stratigraphy exposed in the headwall of the larger thaw slump has been described in Chapter 3 (p. 68) and by Burn *et al.* (1986). The volumetric ice contents of the various cryotextural units are shown in Figure 5.1. The ice above the thaw unconformity is Recent and formed in the last 7 to 8 ka. This ice comprises 9 cm in the ice-rich zone at the base of the present active layer, and over 80 cm in cryotextural unit C (see Figure 5.2).

The oxygen-isotope ratios of ice in unit C (see Figure 3.7, p. 70) indicate that original glaciolacustrine water, now preserved in unit A, has been mixed with younger water of higher $\delta^{18}O$. Further, since the $\delta^{18}O$ values in cryotextural unit C increase upward, we may assume that the young water component in the ground ice also increases. We should expect the least mixing at the base of cryotextural unit C, which was, presumably, thawed

Figure 5.2. Ground ice stratigraphy exposed in an upper section of the smaller thaw-slump headwall, Mayo study site. 16 June 1984.

for a relatively short period both annually and over geological time, and was probably saturated with water melted from unit A ice.

The ground ice in cryotextural unit C comprises lenses 1 - 2 cm thick. Ice lens thicknesses similar to these were found at the base of the present active layer at sites 2 and 7, which are relatively dry. Thicker ice lenses were observed below the active layer at wetter sites (e.g. 1 and 5). Assuming wet conditions in the paleoactive layer, these observations suggest that the lenses in this unit were formed as the permafrost table rose relatively rapidly. Similarly, the relatively thick ice-rich zone at the base of the present active layer indicates stable conditions during its period of formation.

The aggradational ice just below the active layer and in unit C was probably incorporated into permafrost by the processes described earlier. It is unlikely that cracks or fissures were responsible for transporting the enormous amount of excess water into the unit. Between June 1982 and October 1985 the thaw slump retreated 45 m, while the dimensions of unit C remained approximately $1.2 \times 80 \text{ m}^2$. The volume of ice exposed in the unit was on the order of 2000 m^3 , under a surface area of 3600 m^2 .

The 90 cm of ice at the base of the active layer and in unit C indicates a minimum mean rate of water incorporation into permafrost over the past 7 to 8 ka of between 0.1 and 0.2 mm year^{-1} at this site. The ice considered here has accumulated over several millennia, during which local short-term disturbances,

e.g. forest fires, may have led to a deeper active layer and some desiccation of the permafrost, as at sites 1 and 5 in this study. As a result, the present ice content of these units is less than or equal to the total water incorporated into permafrost since the development of the thaw unconformity. Based upon this, the 9 cm of ice immediately below the active layer suggest that active-layer depth in the Mayo area has been relatively stable over the past 1 ka.

The rate of ice accumulation compares well with the results presented in Chapter 4. 0.1 mm is an order-of-magnitude lower than the neutron probe resolution for a 15 cm interval, and is a minimum estimate of the mean rate of permafrost aggradation in the region under stable environmental conditions. It is important to note that the mean minimum rate of permafrost aggradation during the formation of unit C was somewhat greater, perhaps about 0.2 mm year⁻¹, due to the incorporation of soil into the unit as the permafrost table rose.

5.3 Summary.

The investigations reported in this chapter provide further support for the Cheng hypothesis. Calculations show that transport by mass flow, rather than molecular diffusion, has been the dominant process leading to present concentrations of tritium in permafrost. The presence of tritium in the uppermost layers of permafrost indicates a flux divergence between infiltration in summer or fall and desiccation in winter.

The form and volume of segregated ice in cryotextural unit C of the thaw-slump exposure and in the ice-rich zone below the active layer indicate that the processes discussed in this thesis have continued over the Holocene period. It is reasonable to suggest, therefore, that the Cheng hypothesis is a feature of the mass balance of permafrost terrain, and that permafrost acts as a sink for atmospheric waters. Despite the slow rate at which these processes operate, the immense volume of ground ice contained in unit C testifies to their profound hydrogeological significance.

CHAPTER 6

CONCLUSION

6.1 Purpose and activities.

Experiments conducted to evaluate the hypothesis proposed by Cheng (1982) for the origin of segregated ice in the upper layers of permafrost have been reported in this thesis. Such ground ice is widespread in permafrost regions (Mackay 1975; Pollard and French 1980; Outcalt 1982). Cheng's propositions suggest that an annual flux divergence between temperature-induced water flow downward into permafrost in summer and fall, and upward out of permafrost in winter, leads to net water accumulation in the upper centimetres of permafrost. As a result, permafrost is a sink for atmospheric water.

The experiments documented in the thesis were conducted near Mayo, Yukon Territory, between August 1983 and October 1985. The thermal and hydrologic regimes of the soil at eight sites were monitored simultaneously. Six of the sites are located in ice-rich glaciolacustrine sediments, and the other two are in Recent alluvial deposits. Soil temperatures were measured at 15 cm intervals in the surface 2 m of the ground, and at the same time, the hydrologic regime of each site was examined via neutron moderation and with lysimeters. Telescoping heave tubes and magnets were used to measure frost heave. The processes were also

simulated in 2 m long PVC tubes filled with supersaturated soil, which were placed in the ground in August 1983 and extracted at intervals over the following two years.

Cheng's (1982) hypothesis divides the hydrological year into four seasons: snowmelt, summer, freeze-back and winter. These components of the hydrological regime have been documented and integrated to assess the annual water balance of permafrost.

In addition, the tritium content of permafrost in the study area has been investigated to assess water infiltration into permafrost since the atmospheric atomic weapons tests of the late 1950s and early 1960s. Two plots were irrigated with tritiated water in 1983; the migration of this tritium into permafrost was examined at these plots in order to identify the operative process.

Finally, the ground ice stratigraphy exposed in the headwalls of two thaw slumps in the study area has been examined for evidence relating to the hypothesis on a geological time scale.

The experimental work has not only provided an indication of the rate of water movement into permafrost, but also considerable information about the hydrological properties of frozen soils in the study area.

6.2 Results.

Measurements of ground temperatures at the study sites indicated a diversity of thermal regimes. The variation in winter temperature from site to site was principally caused by dif-

ferences in the snowcover and geomorphological setting of each site. The wettest conditions were observed in recently-drained lake sediments, and the coldest at an exposed site, with thin snowcover. Summer temperature variations were associated with the thickness of the surface organic layer (cf. Riseborough 1985) and soil moisture conditions. There was relatively little difference in the temperature gradients recorded at the sites. However, while the temperature gradient determines the direction of water movement, the flow rate is governed by the temperature-dependent hydraulic conductivity.

Lysimeter measurements of evaporation and precipitation indicate that most summer precipitation is absorbed by the surface organic layer and utilized by the plants in this horizon. As a result, active-layer recharge only occurs during snowmelt and intense summer rainfall, when the organic layer is saturated. Changes in the moisture content of mineral soil are independent of the hydrodynamics of the organic horizon. Evapotranspiration by trees and shrubs and downward water movement into frozen ground contribute to summer desiccation of the active layer. As described by Rutter (1975), evaporation from the ground surface accounts for less than 20 percent of total evaporation from the forest.

Downward water movement into frozen ground, accompanied by frost heave, was observed throughout the summer (cf. Mackay 1983a). Up to 1.5 cm of downward flow were detected at the wettest site, where conditions were optimal. Water migration into

permafrost was observed at the same site from mid-July, but at others only after mid-September. Determinations of the hydraulic conductivity of the soils at temperatures below 0°C , based on these observations, compare well with published values ($10^{-11} \text{ m s}^{-1}$ between 0 and -0.6°C , $10^{-12} \text{ m s}^{-1}$ between -0.6 and -1.5°C ; Yoneyama *et al.* 1983; Smith 1985a).

Soil moisture redistribution during freeze-back followed the pattern of enrichment near the surface and desiccation lower in the active layer (Parmuzina 1978; Mackay 1980, 1983a). Some water accumulation in permafrost was also observed at most sites during freeze-back. The upward water movement in the active layer led to segregational frost heave at several sites. At five sites, secondary frost heave was detected by the magnets some distance from the location of primary ice segregation. At one site, where almost 12 cm of heave was observed, secondary heave accounted for at least 2.75 cm (23 percent) of the total.

Between the completion of freeze-back and the start of snowmelt, water movement apparently continued through the soil under the positive temperature gradient. Water accumulation in the surface layer of mineral soil was observed at several sites in this period. In addition, up to 3 mm of vapour outflow from the soil were measured in traps placed on the ground surface. However, relatively little desiccation was observed in the soil profile, implying that this water comprised contributions from several layers, including permafrost.

Most soil recharge occurred during snowmelt, when water

content increases due to infiltration were measured at depths of up to 75 cm. Temperature-induced infiltration of water into mineral soil was observed. A small amount of frost heave was observed during snowmelt infiltration.

Over the two years of the study, a net water accumulation in permafrost was observed at several sites, and was accompanied by ground heave. A maximum rate of water accumulation in permafrost of 3.5 mm year^{-1} was measured. All soil analogue tubes recorded moisture redistribution with net water accumulation in permafrost; in these closed, saturated systems, a mean accumulation of approximately 1 cm year^{-1} was recorded. Large segregated ice lenses were observed in these tubes, just below the base of the active layer. By October 1985, some of these lenses were over 1.5 cm thick.

Measurements of the tritium concentration in drill core samples indicate that since the late 1950s and early 1960s, tritium has migrated at least 50 cm into permafrost. Calculations based on the observed uptake of tritium in permafrost at the sites irrigated with tritiated water in September 1983, indicate that this movement is not due to diffusion, but downward moisture migration induced by the ground temperature gradient in summer.

Finally, observations of the ground ice stratigraphy in the study area also support the hypothesis. The stratigraphy in the thaw slump headwalls contains four principal units: (1) the active layer (0 - 90 cm); (2) an ice-rich zone, 90 percent ice by volume (90 - 100 cm); (3) a second cryotextural unit with ice lenses up

to 15 cm x 2 cm in cross-section (100 - 250 cm); and (4) a lower unit of coarser cryotexture, with ice lenses up to 15 cm x 80 cm in cross-section. Burn et al. (1986) have suggested that the break between units 3 and 4, which truncates several ice wedges, represents a thaw unconformity, that formed during the Early Holocene, about 8 ka ago. The 90 cm of ice above this unconformity therefore provide a minimum mean estimate for the rate of water accumulation in permafrost of 0.1 mm year^{-1} .

6.3 Conclusions.

The principal conclusions of this study may be summarized as follows:

(1) The results of the experiments support the Cheng hypothesis for the origin of aggradational ice in permafrost. As a result of the dependence of soil hydrologic properties on temperature below 0°C , permafrost is a sink for atmospheric water. In the study area, the maximum measured rate of water accumulation in permafrost, and, hence, of permafrost aggradation, is 3.5 mm year^{-1} , or about one percent of total precipitation.

(2) The investigations of tritium infiltration and ground ice stratigraphy indicate that the processes considered are an integral part of the hydrology and hydrogeology of permafrost. The ground ice stratigraphy indicates that over the past 8 ka or so, the mean minimum rate of water accumulation in permafrost has

been 0.1 mm year^{-1}

(3) The results also indicate that the hydrology of mineral soil at the study sites is relatively independent of processes in the surface peat cover. In summer, little evapotranspiration occurs via the peat, but mineral soil in the thawed active layer loses its water to uptake by trees and endohydric plants, or by downward movement into frozen ground.

(4) Frozen soil apparently continues to conduct water, in the presence of a temperature gradient, to temperatures several degrees below 0°C . A small amount of water accumulation in the surface layers of the ground was measured at most sites over the winter. Measurements of vapour outflow from the soil suggest that upward water movement continues throughout the winter.

(5) Segregational frost heave measured during freeze-back indicates that secondary heave (heave on the cold side of the primary ice lens) may account for over 20 percent of total frost heave. Nearly 3 cm of such heave was measured at one site between mid-October 1984 and mid-April 1985. Given this magnitude, it may be of considerable importance to engineering and construction in "warm" permafrost conditions.

(6) Determinations of the hydraulic conductivity of frozen soil based upon the vapour outflow and frost heave measurements are in

agreement with other values in the literature. The determinations suggest that 10^{-12} to $10^{-13} \text{ m s}^{-1}$ may be a conservative estimate of this property for frozen soils.

(7) Data collected during freeze-back and in laboratory tests indicate that the measurement of soil water content by the neutron probe is complicated by the freezing and thawing of soil water. This is not understood fully at present.

(8) The system for measuring frost heave using ring magnets buried at intervals in the soil has been shown to offer considerable promise for long-term use.

As indicated above (p. 191), the processes discussed in this thesis are of considerable importance to the hydrology and hydrogeology of permafrost regions. Their significance to aspects of permafrost geomorphology has also been considered (pp. 54-55). The measurements of secondary frost heave and the determinations of the hydraulic conductivity of frozen soils should be noted by northern geotechnical specialists and may prove to be the results most immediately relevant to the development of Canada's North.

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APPENDIX 1

Soil particle size distribution, sites 1 to 8, Mayo study area.

Site	Depth	% Sand ($>50\ \mu\text{m}$)	% Silt	% Clay ($<2\ \mu\text{m}$)	Soil Type
1	20 - 30	2.0	57.6	40.4	Silty clay
1	70 - 80	8.0	57.1	34.1	Silty clay loam
1	120 - 130	1.3	52.9	45.8	Silty clay
1	170 - 180	1.7	52.5	45.8	Silty clay
2	20 - 30	12.6	58.2	29.2	Silty clay loam
2	70 - 80	55.3	22.2	22.5	Silt loam
2	120 - 130	49.9	47.1	3.0	Sandy loam
2	170 - 180	7.3	74.6	18.1	Silt loam
3	20 - 30	20.9	57.4	21.7	Silt loam
3	70 - 80	6.1	68.5	25.4	Silty clay loam
3	120 - 130	9.8	76.0	14.2	Silt loam
3	170 - 180	4.2	57.8	38.0	Silty clay loam
4	50 - 60	0.8	75.6	23.6	Silt loam
4	100 - 110	1.8	71.4	26.8	Silt loam
4	150 - 160	3.0	77.0	20.0	Silt loam
4	180 - 190	25.0	55.0	20.0	Silt loam

Site	Depth	% Sand ($>50\ \mu\text{m}$)	% Silt	% Clay ($<2\ \mu\text{m}$)	Soil Type
5	40 - 50	2.5	72.4	25.1	Silt loam
5	80 - 90	0.8	74.4	24.8	Silt loam
5	150 - 160	1.2	73.6	25.2	Silt loam
5	190 - 200	1.5	79.7	18.8	Silt loam
6	20 - 30	2.1	76.7	21.2	Silt loam
6	80 - 90	4.0	77.2	18.8	Silt loam
6	140 - 150	0.9	69.0	30.1	Silty clay loam
6	190 - 200	3.2	73.0	23.8	Silt loam
7	30 - 40	7.0	75.0	18.0	Silt loam
7	60 - 70	8.1	77.2	14.7	Silt loam
7	90 - 100	26.2	67.0	6.8	Silt loam
7	140 - 150	24.8	65.2	10.0	Silt loam
8	20 - 30	66.5	29.0	4.5	Sandy loam
8	80 - 90	50.4	38.8	10.8	Loam
8	120 - 130	34.2	57.2	8.6	Silt loam
8	160 - 170	11.9	77.9	10.2	Silt loam

Soil analogue tubes:

1	19.34	64.83	15.83	Silt loam
2	18.6	61.2	20.2	Silt loam

APPENDIX 2

Investigation of the neutron probe measurement of soil water content during freezing and thawing.

The laboratory investigation of the changes in neutron probe measurements of soil water content during freezing and thawing was conducted at Carleton University in April and May 1986. Measurements were made of changes in r (the neutron moderation ratio proportional to soil water content) for two soils and water at various temperatures.

The soils were placed in a plastic bucket, 29 cm diameter x 37 cm deep. The bucket was filled ~~with layers~~ of soil, 5 cm thick, which were frozen before a subsequent layer was added, in order to minimize ice segregation and moisture redistribution during freezing. The soil surrounded an access tube similar to those used at the field sites. The probe was lowered into the soil so that at the base it was 5 cm above the bottom of the bucket. For all readings the bucket and probe were placed in a controlled-temperature incubator to maintain a constant environment surrounding the probe.

The soil was frozen in a commercial freezer at temperatures ranging from -26 to -8°C . The temperature was monitored with a mercury thermometer, calibrated at the National Research Council, Ottawa. The soil was always frozen overnight at -26°C , and

subsequently warmed. It was allowed to equilibrate overnight between readings at different temperatures.

Ten counts with the neutron probe were made for each determination. The mean of these was divided by the standard count to obtain r . The standard counts varied by less than 2 percent for the duration of the tests. The results of the tests are presented in Tables 3.7a and 3.7b (p. 104), and discussed in the main body of the text following p.103.

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