HEAT AND MOISTURE EXCHANGE IN A PERMAFROST ACTIVE LAYER, CHURCHILL, MANITOBA

BY

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A Thesis

Submitted to the School of

Graduate Studies in Partial

Fulfillment of the Requirements

for the Degree Master of Science

McMaster University

February, 1980

HEAT AND MOISTURE EXCHANGE
IN A PERMAFROST ACTIVE
LAYER, CHURCHILL, MANITOBA

MASTER OF SCIENCE (1979)
(Geography)

MCMASTER UNIVERSITY
Hamilton, Ontario

TITLE: Heat and Moisture Exchange in a Permafrost Active Layer, Churchill, Manitoba

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NUMBER OF PAGES: vii, 62

ABSTRACT

This thesis examines heat and moisture exchanges in the active layer of wet tundra soils during freeze-back and thaw. Measurements of rainfall, snow depth, air and soil temperature, soil moisture, and frost heave were recorded daily. Net radiation, soil heat exchange from soil solids and soil moisture, and soil latent heat exchange were calculated. It was determined that soil moisture is the most significant factor influencing soil heat exchange. A prolonged zero curtain effect of 4 to 6 weeks was related to high soil moisture levels. Soil heat exchange as a percentage of net radiation was 10% during snowmelt, decreasing to 3.5% by mid-summer. The latent heat released upon freezing or thawing of soil moisture can increase heat exchange as a percentage of net radiation to 258% during freeze-back. The calculated frost heave of 3 cm indicates that soil moisture was transferred to the freezing front to produce the measured 4.35 cm of heave. These results apply to wet tundra soils with a constant soil moisture.

ACKNOWLEDGEMENTS

I would like to thank Dr. W.R. Rouse for his continual assistance, support, and enthusiasm throughout this study, and Mr. Peter F. Mills for his invaluable assistance in the field and with the instrumentation. I also wish to thank Mr. Gary Miller for the help he gave during the field seasons and for the encouragement and support I received. Mr. Bill Erickson of the Churchill Northern Studies Center also offered much assistance. This study was conducted under grants from the National Science and Engineering Research Council, the National Geographic Society, and the Department of Indian and Northern Affairs.

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CHAPTER 1

INTRODUCTION

This thesis examines a wet tundra surface and the active layer heat and moisture exchanges during freeze-back and thaw. Soil temperature and moisture regimes are influenced by such factors as amount of organic material, the presence of a surface peat layer, soil type, vegetation and snow cover, soil moisture, and the radiation balance. It is the relative importance of such factors with respect to freeze-thaw cycles that is studied.

The chosen field site is near Churchill, Manitoba in the continuous permafrost zone of the open tundra. The Churchill region was selected because it is easily accessible and is in a unique geographical environment. Located at the mouth of the Churchill River (58°45'N, 94°04'W), the wide range of environments including tundra, taiga, and tidal, reflects the combined influence of Hudson Bay and variations in local drainage conditions. The research sites are located on very poorly-drained marine clay beds deposited during an invasion of the Tyrell Sea.

The vegetation is typical of open tundra in the Hudson Bay Lowlands, consisting of lichens, mosses, low shrubs, and isolated, stunted tamarack and black spruce trees. The immediate area contains numerous ephemeral and perennial ponds, peat tussocks, and frost mounds. The two best developed frost mounds were selected as study sites.

Detailed measurements and calculations of rainfall, snowdepth, air and soil temperature, soil moisture, frost heave, incoming and reflected solar radiation, outgoing longwave and net radiation, albedo, soil heat exchange from soil solids and latent heat exchange were made at each of the sites from August 6 to November 24, 1977 and May 4 to September 13, 1978. These give some insight into the moisture and energy exchange processes in an ice-rich active layer.

CHAPTER II

BACKGROUND AND THEORY

Background

Permafrost is characterized by a particular thermal regime whereby soil temperatures remain at or below $0^{o}C$ for a minimum of 2 years. The thermal regime and active layer development are very sensitive to changes at the soil-atmosphere interface. Consequently, man's activities can have a pronounced effect on the equilibriumof permafrost environments. Much of the Canadian Arctic and sub-Arctic is composed of wet tundra clay soils with well developed organic and vegetation layers preserving the particular thermal regime. As Arctic resource exploration and development continues an understanding of the factors influencing thermal regimes and heat exchange becomesincreasingly important.

There are two comprehensive review papers into permafrost research. Brown and Péwé (1973) collate research done in the decade 1963-1973, and Gold and Lachenbruch (1973) discuss thermal conditions in permafrost, summarizing research till that time.

The relationship between permafrost and vegetation has been studied in Canada by Brown and Williams (1972), Nicholson (1978), and Scott (1964), in Russia by Balobaev (1964), and Tyrtikov (1959), and modelled by Ng and Miller (1977). The relationship between permafrost and snow cover has been examined in a number of studies at Fairbanks, Alaska, and Schefferville, Quebec, as is described by Brown and Péwé (1973). Investigations of moisture regimes and movement in permafrost have been examined by Brown (1969), and Martynov (1959).

Comprehensive studies of tundra microclimate energy balances, and active layer regimes have been conducted at Barrow, Alaska by Kelley and Weaver (1969), Weller et al., (1972), Outcalt et al., (1972), Weller and Holmgren (1976), and McGaw et al., (1978). Lewis and Callaghan (1976) conducted energy balance research at an unspecified Arctic site.

Much of the more recent work deals with the modelling of tundra microclimate and active layer regimes, for example, Outcalt (1972), Outcalt et al., (1975), Smith (1975), and Ng and Miller (1975).

Investigation of permafrost in the peat lands and the Hudson Bay Lowlands was initiated in the early 1960's by Brown, for example Brown and Williams (1972), and studies comparing soil temperature and active layer depth between different sites in the Keewatin continue today (Brown, 1978).

In the early 1950's Brown Beckel (1957) studied thermal regimes in the active layer and the effects of snow and vegetation cover at several sites in the Churchill vicinity. Soil temperature was recorded every 2 to 5 weeks, and though no measurements of soil moisture were taken, various wet and dry sites were compared. His temperature measurements did not extend into the permafrost table.

Theory

Heat and moisture exchanges across a permafrost soil surface are driven by solar radiation and regulated by boundary conditions at the active layer base and soil-atmosphere interface. Interface boundary conditions include vegetative cover and snow cover, atmospheric parameters such as temperature, humidity, and windspeed, and such soil factors as thermal and hydraulic conductivity, moisture content, temperature and soil type. The amount of energy gained or lost by the soil is dependant primarily on the magnitude of energy derived from solar radiation and overlying air masses. The transfer of this energy within the soil is dependant on the particular soil profile characteristics. The transfer will always be regulated by the soil layer with the lowest hydraulic or thermal conductivity. All factors are closely interrelated.

Heat stored in the soil is largely transferred by conduction although it can be exchanged by convection and radiation in the air filled portions of soil pores. The rate of soil heat exchange is described in the Fourier equation

$$G = -\lambda \frac{\Delta T}{\Delta z} \qquad (1)$$

where: λ = thermal conductivity
T = soil temperature
z = soil depth

A positive heat flow indicates heat flow away from the surface into the soil. Thermal conductivity increases with increasing soil moisture and increases suddenly in frozen soils.

The heat flux ΔG in a given layer Δz for a given time period Δt is influenced by the heat capacity C such that

$$\Delta G = C \frac{\Delta T}{\Delta t} \Delta z \tag{2}$$

Heat capacity can be calculated according to

$$C = x_s C_s + x_w C_w + x_a C_a \tag{3}$$

where x_s, x_w, and x_a are the volume fractions of soil solids, water and air respectively, and C_s, C_w and C_a the heat capacities of the same elements (Sellers, 1965). The heat capacity of soil solids C_s can be subdivided into

$$C_s = \frac{x_m C_m + x_o C_o}{x_s} \tag{4}$$

where the subscripts m and o designate mineral and organic fractions respectively (Sellers, 1965).

DeVries (1963) finds that, on the average, $C_m = 1.92 \times 10^6 J\ m^{-3}\ {}^oK^{-1}$, $C_o = 2.51 \times 10^6\ J\ m^{-3}\ {}^oK^{-1}$ at 10^oC, and since the heat capacity of air, $1.25 \times 10^3 J\ m^{-3}\ {}^oK^{-1}$, is very small it can be ignored, allowing the above equations to be combined into

$$C = x_s C_s + x_w C_w \tag{5}$$

such that

$$C = 1.92 \times 10^6\ x_m + 2.51 \times 10^6\ x_o + 4.19 \times 10^6\ x_w \tag{6}$$

for liquid soil water, and

$$C = 1.92 \times 10^6\ x_m + 2.51 \times 10^6\ x_o + 1.92 \times 10^6\ x_i \tag{7}$$

for frozen soil water. The heat capacity of water decreases by more than half upon freezing.

At lower moisture contents, heat capacity is less sensitive than conductivity to changing soil moisture content. The rate of heat exchange for a given soil profile peaks at a moisture content of 8 to 20% by volume.

It is in this range that conductivity increases much more rapidly with increasing moisture content than does the heat capacity.

Substituting for heat capacity in eq. (2) gives,

$$\Delta G = (1.92 \times 10^6 x_m + 2.51 \times 10^6 x_o + 4.19 \times 10^6 x_w) \frac{\Delta T}{\Delta t} \Delta z \quad (8)$$

in the thawed state, and

$$\Delta G = (1.92 \times 10^6 x_m + 2.51 \times 10^6 x_o + 1.92 \times 10^6 x_i) \frac{\Delta T}{\Delta t} \Delta z \quad (9)$$

for frozen soils. Using eqs. (8) and (9), the amount of heat gained or lost can be calculated. In soils undergoing a phase change between freezing and thawing, the exchange of latent heat G_L can be calculated as

$$G_L = \frac{\Delta z}{\Delta t} (xw) L, \quad (10)$$

where: L = latent heat of fusion
$= 3.34 \times 10^8$ J m^{-3}.

The total amout of heat exchanged by a soil layer can be obtained by summing eqs. (8) and (10), or (9) and (10).

In soils experiencing a phase change, the latent heat of fusion can represent 95 to 98% of the total heat exchange (Lewis and Callaghan, 1976). The latent heat released by freezing of wet soils can delay complete freeze-back until after mid-winter and contributes to the occurrence of minimum annual soil temperature many weeks after the minimum annual air temperature. The combined effect of an insulating snowcover and release of latent heat decreases temperature fluctuations with depth and helps maintain fairly high soil surface temperatures throughout freeze-back. The thawed portion of the active layer can act as a heat source when freezing does occur and snow cover reduces heat exchange with the atmosphere (Goodrich, 1978)

The presence of water is significant even in the absence of phase changes. Soil moisture amount is closely related to soil type, other factors being equal. Since clay soils have lower porosities and permeabilities, and greater particle radii of curvature than sandy soils, they can retain more moisture at a given suction. Bare clay and peat soils are similar in that both have low seepage rates and minimal movement between layers due to their strong water retentiveness (Tyrtikov, 1959). Kudryavtsev (1959) states that swampy permafrost areas often have mean annual soil temperatures 0.5 to 1.0°C lower than the better drained areas. He suggests this is due in part to greater quantities of heat energy being used in evaporation, which reduces the amount of energy available for soil heating. A thick, insulating snow cover can reverse this, increasing mean annual soil temperature. As previously stated, increasing moisture content in clay soils will inhibit the active layer development because the increased thermal conductivity becomes offset by increased heat capacities (Ng and Miller, 1977 and Kudryavtsev, 1959).

Heat exchange between soil and atmosphere is controlled by the type and condition of surface cover, especially snow and vegetation which protect the surface from wind, regulate moisture exchanges and increase resistance to convective heat loss. Mosses and lichens are the most effective insulators and can impose very steep temperature gradients within the vegetative layer, often causing high surface temperatures. Tyrtikov (1959) suggests that high surface temperatures are accentuated by decreased evaporative cooling at the surface of the vegetation mat due to high resistance to moisture movement. The low thermal conductivity of these plants also decreases soil temperature fluctuations. Moss and lichen mats have a large moisture holding capacity and well-developed hygroscopic properties. Kudryavtsev (1959) states that the moisture holding capacity of sphagnum moss is 1300 to 5000% relative to dry weight and that of hypnum moss 360%. The moisture content of a peat layer can be 5 to 10 times that of mineral soil (Tyrtikov, 1959).

Table 1 shows dry peat to be one of the most effective insulators by virtue of the low heat capacity, conductivity and diffusivity. Snow at $0^{o}C$ and at the same density as dry peat has a greater specific heat and slightly higher heat capacity, which are offset by a much greater conductivity. The conductivity of snow is directly proportional to the square of its density (Tyrtikov, 1959). An increase in the moisture content of a peat layer can increase the conductivity up to 18 times, and if subsequently frozen up to 36 times (Tyrtikov, 1959). The peat layer, if wet, is, therefore, a better conductor in winter, facilitating winter cooling and slowing summer heating. In a soil profile composed of mineral, peat and vegetation layers, and overlain by snow, heat exchange is regulated by the layer with the lowest conductivity.

Ng and Miller (1977) studied the sensitivity of active layer depth of soils at Barrow, Alaska. Thaw depth was most sensitive in the first place to changes in organic layer conductivity and in second place to changes in organic layer depth. Russian studies demonstrate that removal of the moss-lichen mat can increase thaw depth 2 to 3 times, while removal of the peat layer increases it 1.5 to 2.5 times (Tyrtikov, 1959). The organic and vegetation layers are, therefore, the most important factors influencing soil thermal regimes (Brown and Williams, 1972 and Tyrtikov, 1959). Snow cover also affects thaw depth by decreasing early winter heat loss. However, additional energy is required to melt the snow, delaying thaw and contributing to a shallower active layer.

As previously stated, the main energy source for heat and moisture exchange is radiant energy. The radiation balance equation

$$Q^* = K\downarrow (1 - \alpha) + L \quad (1 - \varepsilon) - (\varepsilon\sigma \, T_s^{\,4}) \tag{11}$$

where: T_s = surface temperature
ε = infrared emissivity of the surface
σ = Stephan-Boltzman constant
$K\downarrow$ = incoming shortwave radiation
$L\downarrow$ = incoming longwave radiation
α = surface albedo

gives the net amount of energy available at a surface in the absence

Table 1 Thermal Relationships of Various Surface Covers and Soil Types

Surface	Density ($kg\ m^{-3} \times 10^3$)	Specific Heat ($J\ kg^{-1}\ {}^oK^{-1} \times 10^3$)	Heat Capacity ($MJ\ m^{-3}\ {}^oK^{-1}$)	Thermal Conductivity ($Wm^{-1}\ {}^oK^{-1}$)	Thermal Diffusivity ($m^2 s^{-1} \times 10^{-7}$)
Soils-Clay (dry)	1.8	1.46	2.514	1.220	4.85
Sand (wet)	1.6	1.68	2.514	2.510	9.98
Sand (dry)	1.0	1.26	1.257	1.130	8.99
Peat (dry)	0.5	1.26	0.838	0.062	0.74
Still Air	0.0013	1.01	0.001	0.002	20.00
Water (20°C)	1.0	4.19	4.196	0.587	1.40
Ice (0°C)	0.91	2.10	1.927	2.260	11.70
(-45°C)	0.91	1.76	1.592	2.810	17.70
Snow (0°C)	0.1	2.11	0.209	0.075	3.59
	0.3	2.11	0.628	0.260	4.14
	0.5	2.11	1.047	0.612	5.85
	0.7	2.11	1.466	1.230	8.39
	0.9	2.11	1.885	2.500	13.30
Snow (-45°C)	0.1	1.78	0.167	0.075	4.49
	0.3	1.78	0.544	0.260	4.78
	0.5	1.78	0.879	0.612	6.96
	0.7	1.78	1.257	1.230	9.79
	0.9	1.78	1.592	2.250	14.10

(Geiger, 1965; List, 1958; Sellers, 1965) of advection. Net radiation is strongly influenced by surface albedo. For a given vegetation type and solar zenith angle, surface albedo is influenced by the amount of surface water and whether it is liquid or ice. Tundra albedos vary from 0.20 for a dry surface to 0.15 for a wet surface immediately after melt. New snow cover can increase it to 0.80 (Weller and Holmgren, 1976). Snowpack ripening and decay cause a rapid albedo decrease as patches of wet tundra appear. The longwave radiation balance is influenced by surface temperatures which can be very high in a moss-lichen mat and by atmospheric temperature and cloud cover.

The net radiant energy can be used for sensible heating H, evaporation LE, and soil heating G so that

$$Q^* = H + LE + G. \tag{12}$$

A comparison of the fluxes throughout the Arctic year shows changes of orders of magnitude, each flux reversing it's direction in winter. Large summer net radiation is due to small albedos and long daylight hours. Q^* is negative during the winter periods when $K\downarrow$ is small and there is high albedo and continuous snow cover.

A study of the microclimate at Barrow, Alaska by Weller and Holmgren (1976) demonstrates that soil heat flux is negative from freeze-back through winter as the ground releases heat to the atmosphere. Sensible and latent heat fluxes are also negative in winter, indicating frost deposition by sublimation and eddy heat flux towards the surface, though the authors state these results could not be confirmed by measurement. The most notable change is in relative orders of magnitude. Absorbed radiation increases an order of magnitude upon snow melt and the latent heat flux by a factor of 40. By the end of snowmelt, 9% of Q* is used for soil heating, 18% for sensible heating, and 73% for evaporation, while in mid-summer the percentages are 2%, 32% and 66% respectively.

On a diurnal basis, the maxima and minima of the sensible and latent heat fluxes usually correspond to those of net radiation, yet McKay and Thurtell (1978), working in Southern Ontario, found that in winter, much of the energy stored in the snowpack came not from net radiation but from warm air masses associated with large scale weather systems. At higher latitudes where incursions of warmer air are less frequent, this effect would be less pronounced.

Much of the higher Arctic and 92% of the Hudson Bay Lowlands is composed of wet surfaces, yet evapotranspiration in high latitudes is more strongly influenced by vegetation forms than in lower latitudes (Rouse, 1973 and Rouse and Kershaw, 1971). Unlike vascular plants, lichens do not transpire. Kershaw and Rouse (1971), Rouse and Kershaw (1973), and Stewart and Rouse (1978) report consistently high soil moisture contents under various burned and lichen surfaces indicating that even on better drained sites evaporation is not limited by soil moisture but rather by xeric vegetation. Stewart and Rouse (1976), observed that it is where there is standing water or freely-available water in shallow lakes and wet tundra that the highest evaporation occurs.

Four phenomena which influence freezing and thawing of permafrost soils are freezing point depression, moisture migration to a freezing front, frost heave and settlement, and the zero curtain effect. As previously discussed, the zero curtain effect occurs when latent heat released upon freezing of soil water delays complete freeze-back. The freezing point of water under pressure is lower than that of free water. The closer to the soil particle surface, the more tightly bound the water and the greater the freezing point depression. There is, therefore, a freezing point depression gradient from free water to pore water to adsorbed moisture films (Jumikis, 1966). Freezing point depression is, therefore, related to particle size and is greater in clays. It also increases with increasing salt concentrations. Tsytovich (1975) determined experimently that freezing can occur at $0^{o}C$ in sand, -0.1 to $-1.2^{o}C$ in plastic clays, and -2.0 to -5.0^{o} for semi-hard and hard clays.

There are many theories pertaining to the mechanism of soil moisture migration to a freezing front, of which the adsorption film theory is the most comprehensive in that it includes moisture movement in both thawed and frozen soils. Soil particles are surrounded by liquid water films of varying thickness and movement is towards particles with thinner films. All moisture is not fixed upon freezing but movement continues at a decreased rate in the more tightly bound water films. The adsorption forces of soil particles are governed by the free surface energy of the particles, and determine both the rate of moisture migration and the amount of frost heave. Moisture migration is also regulated by a soils capillary properties.

In situ pore water expands by a factor of 1.09 on freezing. With the migration of water into a zone and its subsequent freezing, substantial localized expansion will occur and be visible as frost heave. The rates and amounts of frost heave and settlement are dependant on moisture migration, water supply, and particle size and density, while the heave rate also depends on the rate of heat removal and frost penetration (Penner, 1972). The more rapid the frost penetration, the less water attracted to a given soil volume. Heave rate in sand is minimized by decreased capillary forces and smaller moisture flow, whereas in clay, heave is limited by low permeabilities. Maximum heave, therefore, occurs in silts (Scott, 1969). Tsytovich (1975) determined experimentally that an initial temperature gradient of $5^{o}C/cm$ was required to overcome internal resistance to moisture movement in stiff plastic clays. It was also determined that heaving in fine clay can continue to $-10^{o}C$ and sometimes at even colder temperatures as layers of more tightly bound water freeze.

The magnitude of the heave force depends on such soil properties as particle size, cation exchange capacity, compressibility of lower horizons and overburden pressure.

CHAPTER III

METHODS

Site Description

The research area is situated approximately 6 km east of Churchill, Manitoba in a hummock bog terrain (Figure 1). To the southeast is an outlier of the black spruce forest, to the northeast a Pre-Cambrian ridge approximately 28 m high, and treeless tundra in all other directions. There are numerous ponds and small lakes in the immediate area and the terrain is saturated, often covered with standing water due to poor drainage (Figures 2 and 3). Two frost mounds (large mounds with a mineral soil core) approximately 7 m in diameter and 35 m apart were selected from the many smaller peat tussocks. Clay soils were selected since drier, sandy soils undergo little frost heave and have a smaller total soil heat exchange. No silt soils were found in the vicinity.

The surfaces of the frost mounds are 1 m above the surrounding terrain. Short grasses and marsh sedge comprise the vegetation of the surrounding bog. On the drier, elevated peat tussocks Labrador tea (Ledum groenlandicum), some lichen, and the odd stunted black spruce (Picea mariana) are found. The vegetation

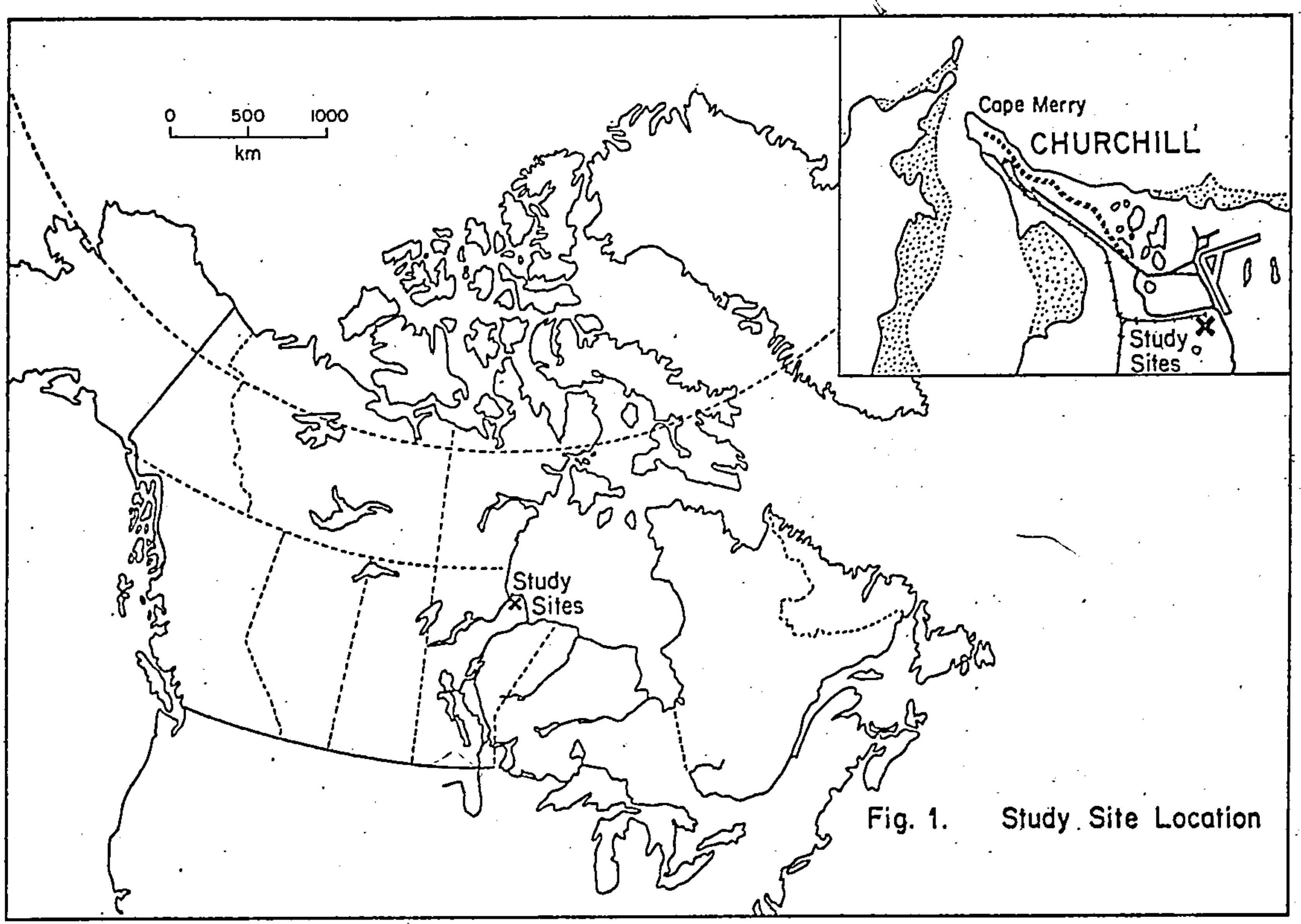


Fig. 1. Study Site Location

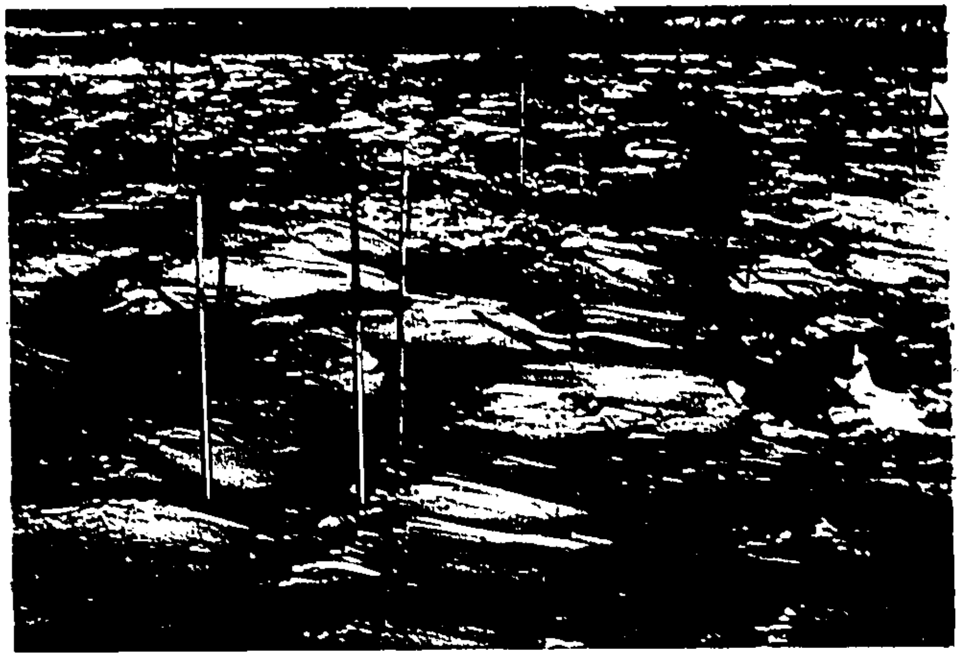


Figure 2a Study Sites, November 1977

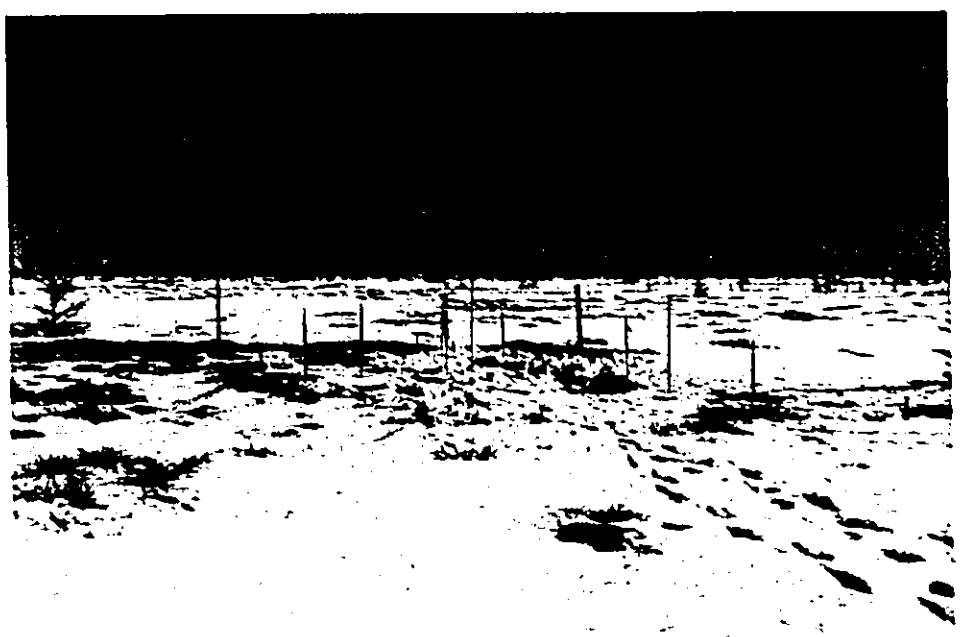


Figure 2b Study Sites, early May 1978

Fig. 3a Study Sites; late May 1978

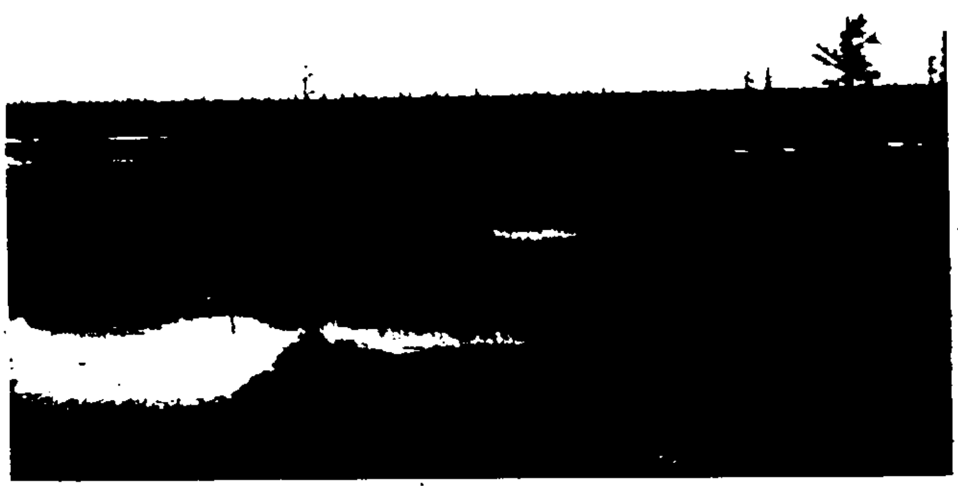


Fig. 3b Study Sites, September, 1978

on the mounds is composed mainly of moss, including heather and sphagnum) and lichen (Cladonia spp., Cetraria spp.), with bear berry (Arctostaphylos rubra), and some grasses. Blueberries (Vaccinium spp.) and dwarf willow (Salix spp.) grow around the perimeter.

Soil pit analyses of the frost mounds indicated the following. The thickness of the organic layer varied from 10 to 15 cm. The brownish colored root zone extended to depths of 5 cm while the remainder of the organic layer was composed of black, dense peat in varying stages of decomposition. At the base of the organic layer was a 2 cm band of pebbles mixed with clay. In some profiles there were also larger frost shattered limestone plates aligned in shingle fashion parallel to the surface. A heavy grey clay comprised the remainder of the profile, broken by a very thin, discontinuous gravel band at 55 cm. The active layer ranged in depth from 60 to 125 cm and tended to be deeper at the perimeter of the mounds.

The entire profile was saturated at all times. While digging soil pits, infilling by water seepage was most rapid below 30 to 40 cm and in many cases it was impossible to excavate to the permafrost layer. After 24 hours the standing water in the soil pit was more than 15 cm deep.

Bulk density and the percentage of organic matter with depth were determined by sampling and are listed in Table 2. The average bulk density of clay from the lower horizon was 1880 kg m^{-3}. Since the average bulk density for clay averages 1000 to 1600 kg m^{-3}, these soils are dense and compacted and have a low porosity (Buckman and Brady, 1974). The mineral horizon had a fine clay texture and when wet was plastic. The material became hard and cloddy when dried. The bulk density of the organic layer is near average.

Instrumentation and Recording

The instrumentation was replicated at the two adjacent frost mounds and consisted of the following installations on each mound: 6 neutron probe access tubes, 3 temperature profiles using thermistors, 1 air temperature thermistor, 3 frost heave tubes, and 3 frost depth tubes. The equipment was installed during August 1977. The continuous measurement period for freeze-back was September 1 to November 31, 1977, though some data were collected in early August. Thaw data were recorded from May 1 to September 13, 1978. Most variables were measured once a day in the early afternoon, weather conditions permitting.

The following measurements were made: soil temperature, net radiation, incoming and reflected solar radiation, and outgoing longwave radiation. These were all monitored through a manual stepping switch and the values recorded on a Keithley digital potentiometer. Soil temperature was measured at 0, 5, 10, 20, 30 cm and further 10 cm increments downward through the active layer such that the deepest thermister was originally situated in the permafrost table in early August. Air temperature was measured using a shielded thermister at a height of 1 m. Soil moisture readings were taken at 10 cm intervals from 10 cm to a maximum of 105 cm using a Nuclear-Chicago Subsurface Moisture Probe and Scaler. Surface soil moisture was determined gravimetrically.

Rainfall was recorded daily as the average of 3 rain gauges at each site. Snow depth was also measured daily with meter stick and the series of readings averaged. Frost heave was taken as the change in vertical displacement between a buried dowel and a free riding tube. The wooden dowel was anchored 15 to 20 cm in the permafrost table. The lubricated aluminum tube was fitted over the dowel so that its base rested on the permafrost table. Modified frost depth tubes (Rickard and Brown, 1971) were installed to provide a second measurement of frost penetration, but they performed unsatisfactorily and the data were not used.

During selected clear sky days, radiation measurements were made at 15 minute intervals for a limited period. The shortwave

Table 2
Heat Capacity of Soil Solids

Depth (m)	Bulk Density (kg m^{-3} x 10^3)	Percentage of Organic Material	Volume Fraction Solids	Clay	Peat	Heat Capacity (MJm^{-3} $^oK^{-1}$)
0.05	0.20	100.0	0.54	0.00	0.54	1.358
0.10	1.08	14.1	0.55	0.47	0.08	1.106
0.20	1.85	1.2	0.55	0.54	0.01	1.064
0.30	1.81	1.2	0.59	0.58	0.01	1.144
0.40	1.87	1.0	0.59	0.59	0.00	1.135
0.50	1.84	0.8	0.59	0.59	0.00	1.135
0.60	1.88	0.7	0.59	0.59	0.00	1.135
0.70	1.88	0.6	0.59	0.59	0.00	1.135
0.80	1.88	0.6	0.59	0.59	0.00	1.135
0.90	1.88	0.6	0.59	0.59	0.00	1.135
1.00	1.88	0.6	0.59	0.59	0.00	1.135
1.10	1.88	0.6	0.59	0.59	0.00	1.135
1.15	1.88	0.6	0.59	0.59	0.00	1.135
1.25	1.88	0.6	0.59	0.59	0.00	1.135

fluxes were measured using Eppley Pyranometers, the longwave flux with an Eppley Pyrgeometer, and net radiation with a Swissteco Net Radiometer, all of which were monitored on the digital potentiometer. In order to obtain a continuous net radiation record, net radiation as measured at the study sites was regressed on Q^* as recorded at the nearby McMaster Research Station for the 1978 season, and was regressed on $K\downarrow$ as recorded at the Atmospheric Environment Service Upper Air Station for the 1977 season. The regression equations are listed in Table 3.

Heat Exchange Calculations

In order to calculate the heat exchange in the active layer, the heat capacity with depth was calculated using the clay and organic volume fractions shown in Table 2. The clay content in the 0 to 5 cm vegetation and organic layer was assumed to be zero. The heat capacity was calculated from eq. (6) and (7). Time periods of 6 to 23 days were arbitrarily selected and the change in soil temperature of each layer was determined. Heat exchange from soil solids and water was calculated using eq. (8) and (9), and latent heat exchange using eq. (10). It is assumed that all water froze and thawed at $0^{\circ}C$. Though Williams (1972) estimates that neglecting freezing point depression could result in a 20% error in freezing depth calculations, such an omission in these calculations results

Table 3
Regression Equations for
Q^* on $K\downarrow$

Surface Cover & Condition	Regression Equation	(r^2)
1977		
Very Wet Tundra	$Q^* = -142.58 + (.85)K\downarrow$	0.94
Wet Tundra	$Q^* = -163.03 + (1.01)\ K\downarrow$	0.99
Dryer Tundra (Autumn)	$Q^* = -120.53 + (0.9)\ K\downarrow$	0.97
Continuous Snowcover	$Q^* = -96.24 + (0.4)\ K\downarrow$	0.99
1978		
Wet to Very Wet Tundra	$Q^* = -107.6 + (1.3)\ Q^*_m$ +	1.00

+ Q^*_m is net radiation as recorded at the McMaster Research Site only in a time lag in the G term.

The total heat exchange, therefore, is

$$G = G_s + G_w + G_L \tag{17}$$

where a positive exchange indicates soil warming and a negative value, soil cooling.

CHAPER IV

MACROCLIMATE AND PERMAFROST AT CHURCHILL

Macroclimate

Hudson Bay exerts the main influence on summer climate in the Churchill area. Water temperatures rarely exceed 7^{o}C except close to the shore and pack ice persists into July (Thompson, 1968). Pack ice postpones spring warming, cools onshore summer winds producing fog, and is the cause of frequent autumn snowstorms when onshore winds blow over open water. Open leads are associated with steam fog or sea smoke during winter.

Upper air flow over the Bay is counterclockwise and centered on a low pressure vortex over Davis Strait. Though summer airflow is mainly from the northwest the weakened vortex permits warm, moist, southern air masses to pass directly over the Bay. In autumn the vortex intensifies and shifts south, blocking the storm tracks. Low pressure cells enter the region at 3 to 4 day intervals in the fall.

During December, snowfall and windspeed are at an annual maximum and are closely related to wind direction and pack ice conditions. Winter is characterized by very short days, an ice-covered Bay, clear skies, and high pressures. Snowstorms are due to incursions of southern air masses and last up to 2 days, once or twice a month. Blowing snow can limit visibility for days. The minimum average air temperature is reached in February. Churchill has an average of 25 days a year colder than -34.5°C (Thompson, 1968). The presence of the frozen ocean prolongs winter conditions well into May. Snowstorms can occur till mid-June. Small solar radiation fluxes associated with increasing low level cloud in late winter also delay the onset of spring. Churchill harbour is usually open June 12 although the Bay remains frozen to the north (Thompson, 1968).

Summer rainfall reaches a maximum during July and August when the stormtracks of the Westerlies reach furthest north. Precipitation patterns are dependant on wind direction since cold Bay temperatures inhibit shower and thunderstorm development. The Bay also causes steep temperature inversions which can create seafog up to 7 days a month (Thompson, 1968). Cool daytime summer temperatures reflect the combined influence of a long daylight period and the moderating effect of Hudson Bay.

Weather During the Study Period

Table 4 compares the weather during the study period to the 33 year record from the Churchill Airport Weather Office. The 1977 freeze-back period was warmer than normal. As a result, there was more rainfall and less snow than usual. The 1978 thaw period was colder and wetter than normal.

Table 4

Churchill Weather During
the Study Period +

Month		Mean Monthly Daily Temperature (± °C from normal)	Monthly Rainfall	Average Monthly Bright Sunshine Hours % of normal	Snow-fall
1977	A	-1.8	103	77	0
	S	+1.5	51	110	0
	O	+3.5	107	142	36
	N	+0.4	575	133	71
	D	+0.5	3	119	130
1978	J	+3.1	-	153	67
	F	+6.1	-	104	156
	M	-2.4	-	117	73
	A	+1.0	13	99	139
	M	-0.1	529	82	326
	J	-0.8	64	115	79
	J	-1.8	166	91	-
	A	-2.5	172	81	100

+As compared to the 33 year record
at the Churchill Airport Weather Office.

Permafrost at Churchill

Churchill is situated in the continuous permafrost zone where the average thickness is 60 m (Brown, 1978), though the permafrost thins out under Hudson Bay and along the Churchill River. A study by Brown (1978) examined variations in thermal regime and active layer depth at 4 sites in the Churchill vicinity: bedrock, marine clay and till, palsa, and palsa depression. Ground temperatures to a depth of 15 m were recorded monthly or quarterly. The soil temperature regime was found to be strongly affected by vegetation, peat and snow cover, and soil or rock type (Table 5). The bedrock site experienced the greatest annual soil temperature fluctuation and the lowest mean annual soil temperature due to high thermal conductivity. The amplitude of temperature change at the base of the bedrock profile was 3^{o}C compared to 1^{o}C at the other sites. Mean annual ground temperature increased from marine clay to palsa, to palsa depression with increasing snow depth and peat thickness, and decreasing snow density. There was no permafrost at 15 m in the palsa depression, though temperature profiles indicated its' presence at 30 to 40 m. This was attributed to swampy conditions and increased snow cover.

Numerous polygons, coalescing peat plateaux, palsas, and frost shattered rocks in the Churchill area are evidence of permafrost and active freeze-thaw cycles.

Table 5
Air and Ground Temperatures
and Active Layer Depth at
Churchill, Manitoba+

Site	Average Bottom Hole Temperature (°C)**	Thickness of Active Layer (m) 1974	1975	1976
Bedrock	-2.9	7.39	7.68	7.77
Marine Clay and Till	-2.6	0.61	0.94	0.76
Palsa	-0.9	0.46	0.52	0.52
Palsa Depression	+0.4*	0.16*	0.64	0.76

+ mean annual air temperature -7.3°C
* no permafrost, data for seasonal frost penetration
** average depth 15m.

(Brown, 1978)

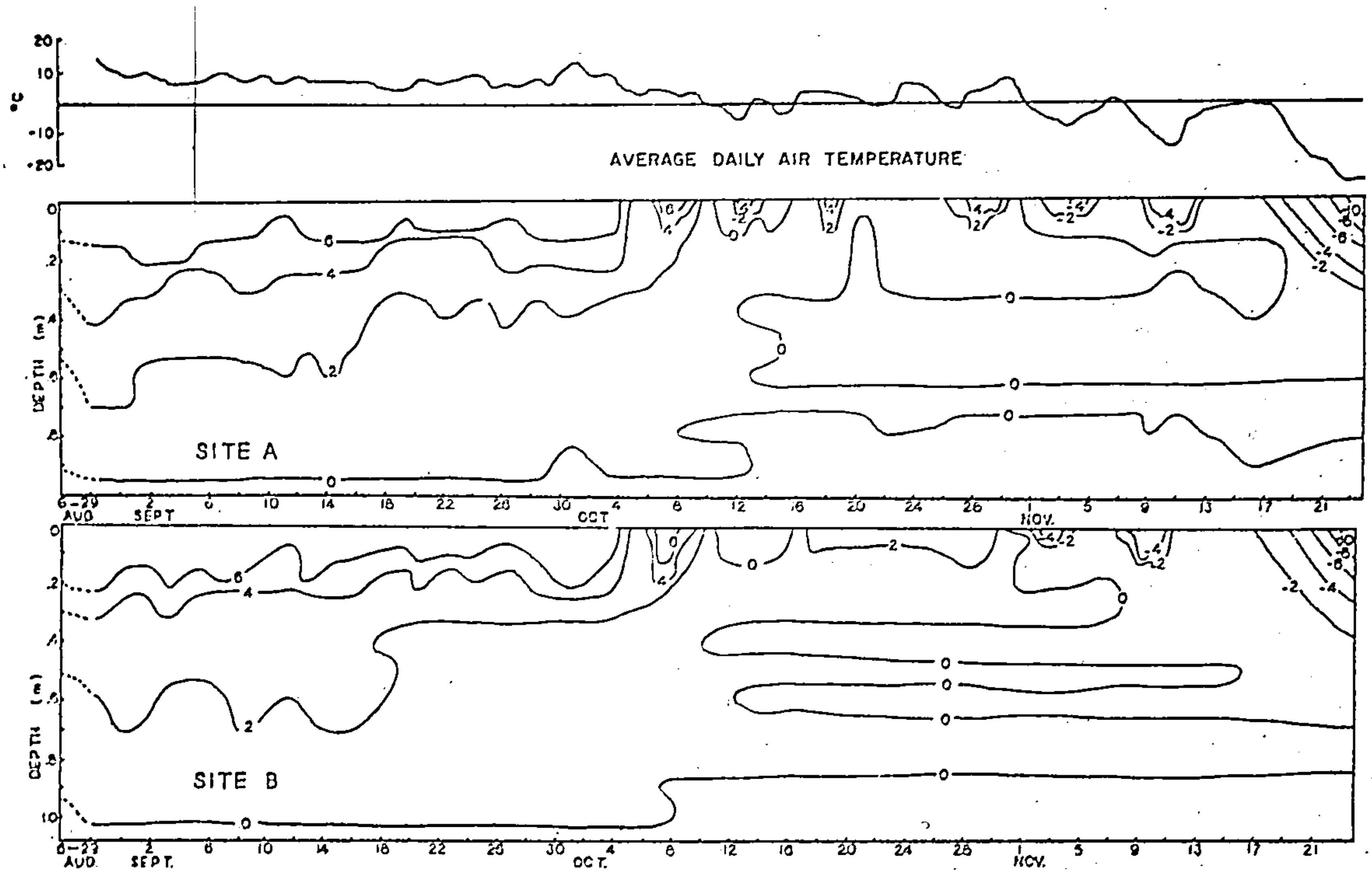
CHAPTER V

THERMAL AND MOISTURE REGIMES DURING FREEZE-BACK AND THAW

Soil Temperature

Figure 4 shows that during freeze-back in 1977 there was upfreezing from the permafrost table, progression of the surface freezing front downward a zero curtain effect, and the development of a frozen layer in mid-profile. Both sites show very similar behaviour. Only the -10^{o}C to 6^{o}C isotherms were graphed. The depth of the permafrost table remained virtually constant from late August through the first week in October when there was rapid upfreezing through 20 cm of soil. There was no further upfreezing until near the end of November and the permafrost table remained separated from the frozen surface layer by a thawed zone. Had measurements continued into the winter the profile would have frozen completely. Upfreezing from the permafrost table has been documented by Brewer (1958), MacKay (1973), and Kelley and Weaver (1969). Drew et. al., (1958) working at Barrow, Alaska found that upfreezing proceeded much faster in drier soils since the high moisture contents, thick organic layer, and snow cover of the wet tundra soils retarded upfreezing. Surface freezing recommenced at the beginning of November and the near surface layers

FIG. 4 SOIL TEMPERATURE 1977

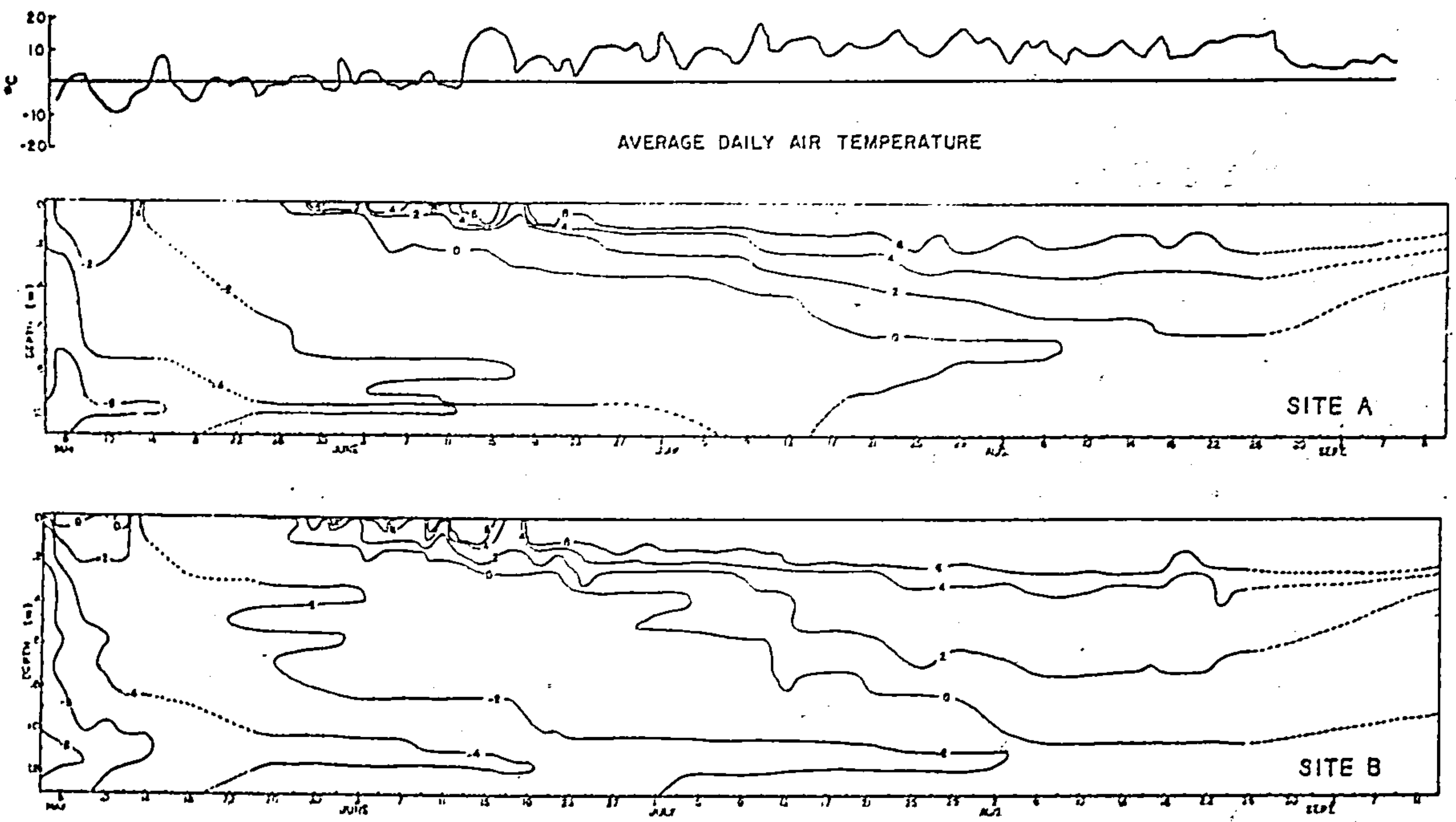


cooled rapidly. The frozen layer in mid-profile merged with the surface freezing front by mid-November.

Temperatures in the mid-profile frozen lens hovered just below freezing until late November, a prolonged zero curtain effect of 4 to 6 weeks which was most obvious below 25 cm. Since a thawed layer from 60 to 80 cm existed at the end of the measurement period, the overall zero curtain effect could have been much longer. Kelley and Weaver (1969) report a zero curtain effect in tundra soils at Barrow, Alaska of 2 weeks in September 1965 and 4 weeks in September 1966. The assumed moisture content was 30% and the active layer depth 60 cm. Lewis and Callaghan (1976) report a 2 to 4 week zero curtain effect for their unspecified Arctic site. The longer zero curtain effect in the present study could be due to a deeper active layer, wetter soils and a very slow freeze-back caused by abnormally high October temperatures. Thaw patterns in 1978 are shown in Figure 5.

Though minimum soil temperatures undoubtedly occured much earlier in the winter, the lowest measured temperatures occured May 5, 1978, the lowest temperature being -9.3^{o}C. It is evident that warming and thawing progressed rapidly from the surface downward. The -4^{o}C isotherm dropped from near the surface to the base of the active layer by the end of May and by early August all soil temperatures were greater than -2^{o}C. The zero curtain effect is quite evident

FIG. 5 SOIL TEMPERATURE 1978



during thaw. Soil temperatures at depth rise to near freezing by mid-June and remain there until mid to late July. A steady downward thaw front progression begins in late May, and merges at 70 cm in early August with the upward moving thaw front which was initiated in mid-July.

Soil immediately above the permafrost table was observed to thaw before the mid-profile section. This was also documented at Churchill by Brown Beckel (1957). Maximum soil temperatures occured at the end of August. By mid-September cooler weather had caused 5 cm of upfreezing. Upfreezing in the unusually warm 1977 season did not begin until the beginning of October. Maximum thaw depth in late August in wet tundra soils was also recorded by Brown Beckel (1957) at Churchill, and Drew et al., (1958) and Outcalt et al., (1975) at Barrow. Figure 6 underestimates the minimum soil temperature curves since minimum annual temperature was not measured. Annual temperature fluctuations decrease with depth and continue on into the permafrost table. Brown Beckel (1957) found that the time between the occurence of mean minimum air temperature and mean minimum soil temperature was greater in wet soils.

During snow melt, temperatures at the study sites increased at an average rate of 0.45^{o}C per day, but this decreased to 0.1^{o}C per day when the weather became cool and overcast during snowmelt (Table 6). Outcalt et al., (1975) suggests that long wave radiation from the base of a low, warm, continuous cloud cover, such as is found at Churchill during snowmelt (Table 7) could contribute markedly to snowpack ripening and decay

FIG. 6 ANNUAL MAXIMUM-MINIMUM

SOIL TEMPERATURE

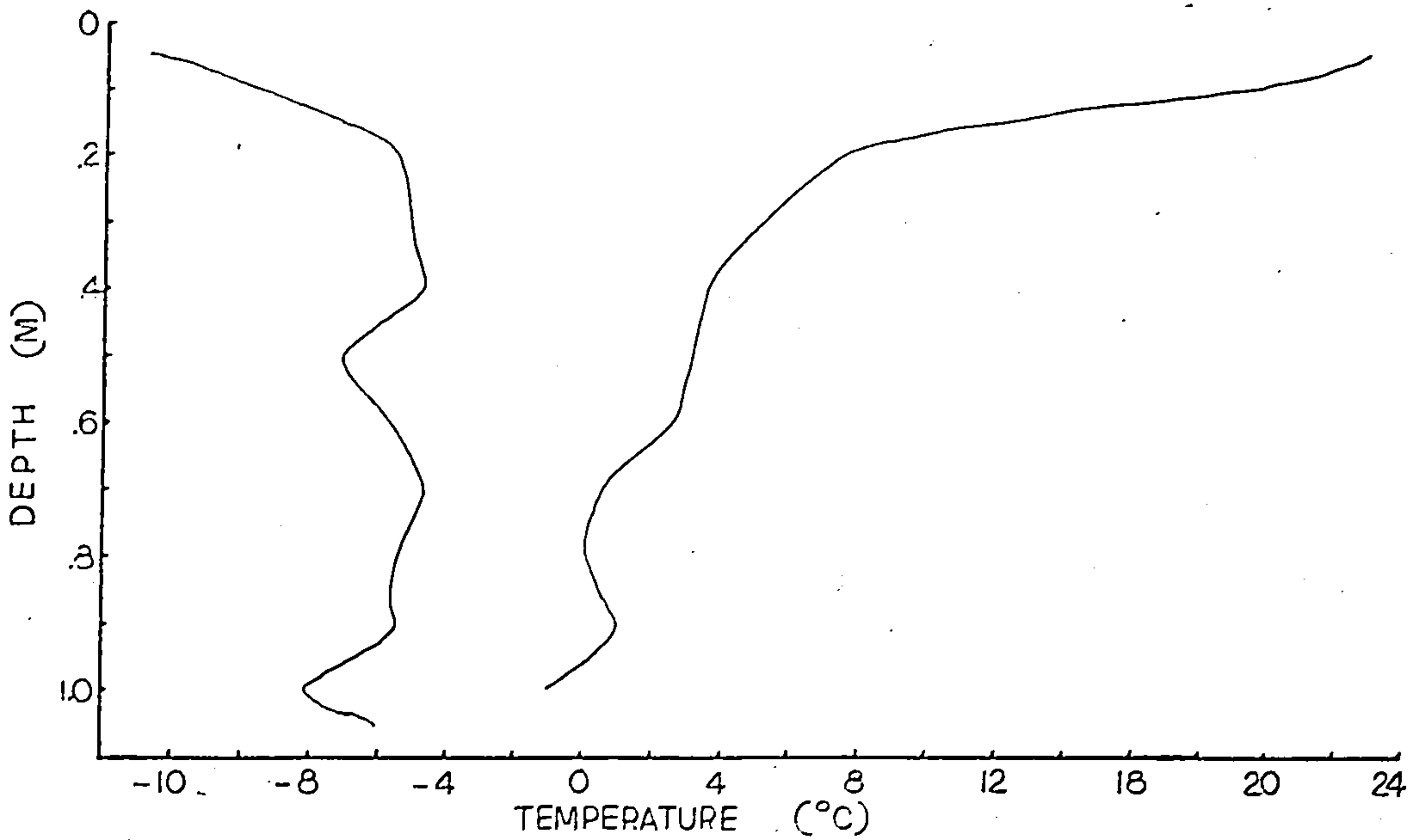


Table 6
Average Daily Soil
Temperature Change (^{o}C)

Date	Site A	Depths (m)		Site B	Depths (m)	
	(.00-.25)	(.25-.45)	(.45-1.05)	(.00-.25)	(.25-.45)	(.45-1.05)
1977						
Aug. 6-29	+0.12	+0.04	+0.02	+0.06	+0.03	+0.01
Aug. 29-Sept. 6	-0.33	-0.13	-0.04	-0.13	-0.11	0.00
Sept. 6-12	+0.10	+0.01	+0.02	+0.18	+0.01	+0.03
Sept. 12-19	-0.70	-0.17	-0.04	-0.85	-0.17	-0.07
Sept. 19-28	+0.70	-0.01	-0.02	+0.69	+0.02	-0.01
Sept. 28-Oct. 8	-0.78	-0.12	-0.03	-0.81	-0.14	-0.04
Oct. 8-18	-0.03	-0.09	-0.06	-0.15	-0.09	-0.06
Oct. 18-29	-0.05	+0.01	+0.02	+0.05	0.00	0.00
Oct. 29-Nov. 8	-0.21	-0.02	-0.02	-0.15	-0.01	0.00
Nov. 8-16	-0.05	+0.03	+0.03	-0.09	-0.01	0.00
Nov. 16-24	-1.08	-0.31	-0.03	-0.96	-0.20	-0.02
Nov. 24-May 4	+0.04	-0.02	-0.04	+0.02	-0.03	-0.04
1978						
May 4-13	+0.22	+0.21	+0.24	+0.20	+0.34	+0.37
May 13-24	+0.03	+0.15	+0.12	+0.05	+0.14	+0.12
May 24-30	+0.15	+0.03	+0.05	+0.17	+0.05	+0.07
May 30-June 8	+0.07	+0.04	+0.04	+0.12	+0.03	+0.04
June 8-17	+0.07	+0.10	0.00	-0.01	+0.05	+0.04
June 17-24	+1.22	+0.01	+0.09	+1.59	+0.19	+0.04
June 24-July 1	+0.33	+0.08	+0.04	-0.52	-0.02	+0.05
July 1-8	-0.68	+0.08	+0.02	-0.26	+0.04	+0.04
July 8-14	+1.22	+0.23	+0.06	+1.52	+0.41	+0.08
July 14-20	-0.28	+0.08	+0.08	-0.37	-0.03	+0.06
July 20-28	+0.31	+0.06	+0.08	+0.13	+0.08	+0.09
July 28-Aug. 3	-0.67	+0.07	+0.08	-0.28	0.00	+0.06
Aug. 3-13	+0.41	-0.03	+0.02	+0.22	-0.01	-0.01
Aug. 13-25	+0.10	+0.08	+0.05	+0.22	+0.09	+0.06
Aug. 25-Sept. 13	-0.21	-0.14	-0.06	-0.26	-0.15	-0.07

Table 7
Average Cloudiness for
Representative Periods+

Period	Dates	Average Cloudiness (Tenths)
Pre-melt	May 1-5	5.0
Snowmelt	May 21-29	8.9
Post-melt	June 7-12	7.1
Mid-summer	July 21-25	5.5
Freeze-back	Nov. 24-28	3.3

+Churchill Airport Weather Office

Figures 7 through 10 plot representative temperature profiles throughout the season. The effects of thermal conductivity on the amplitude of temperature change are evident. During freeze-back when soil water is in the liquid state, conductivity is much smaller than when much of the profile is composed of ice as during thaw. Response to surface heating and cooling is therefore more rapid in frozen soils.

Maximum active layer depths at the study sites average 83 cm and ranged from 60 to 120 cm. Tsytovich (1975) states that average active layer depths for clay soils in the southern Arctic to be 100 to 250 cm, depending on moisture contents. Brown (1978) found active layer depths in marine clay near the study sites to be 61 cm in 1974, 94 cm in 1975, and 76 cm in 1976, while Brown Beckel (1957) found active layer depth in clays at Churchill to range from 0.91 to 2.4 m and in saturated soils, 1.1 to 2.3 m.

As is shown in Table 6, the greatest average daily rate of temperature change occurs near the surface. The periods of greatest warming in late June and mid-July are related to very warm air temperatures. The effect of weather on temperature change is also obvious in September 1977 when the 0 to 25 cm layer cooled, warmed, and re-cooled rapidly, coinciding with the first fall frost, a warm period and further frost and snowfall.

FIG. 7 SOIL TEMPERATURE PROFILES, SITE A, 1977

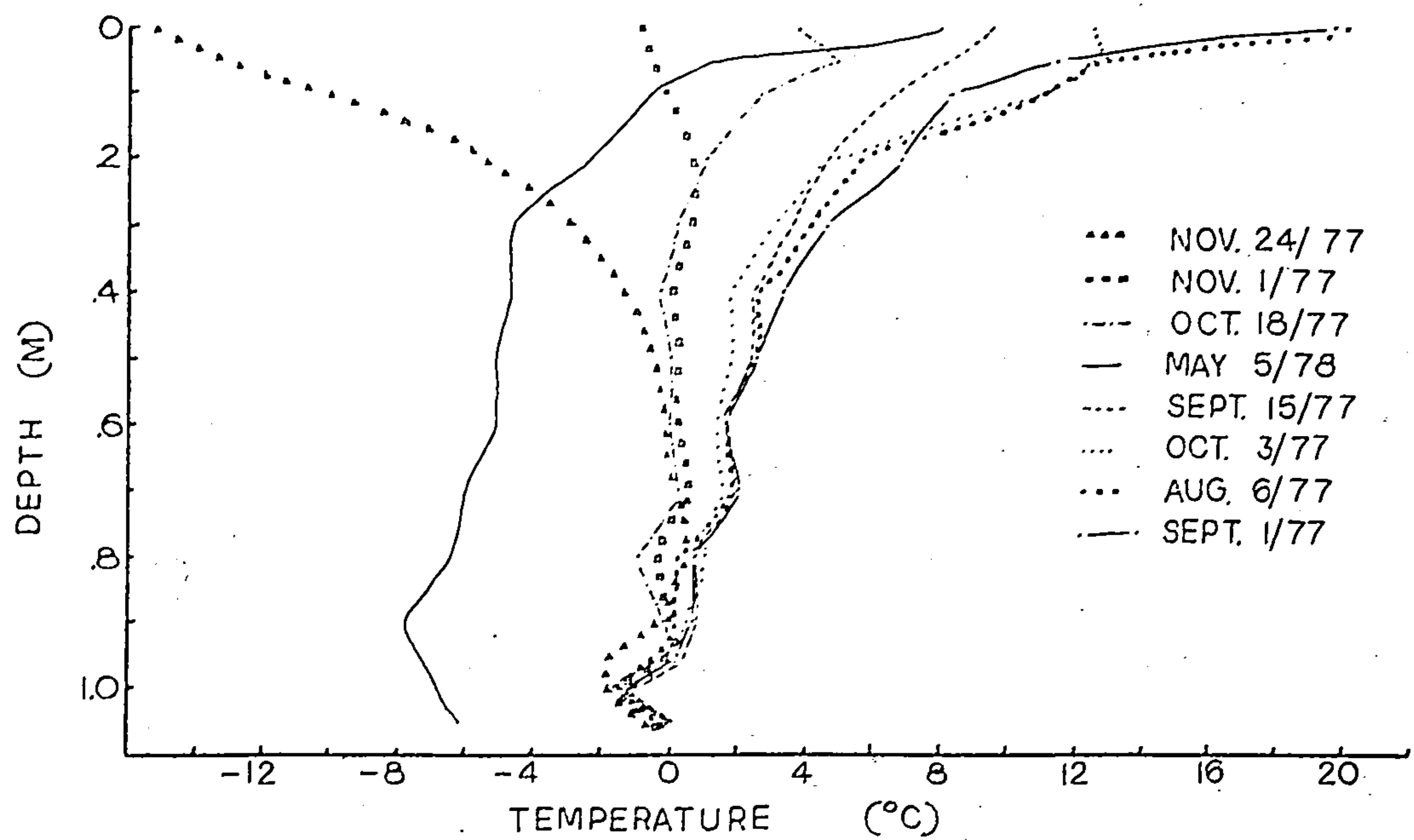


FIG. 8 SOIL TEMPERATURE PROFILES, SITE B, 1977

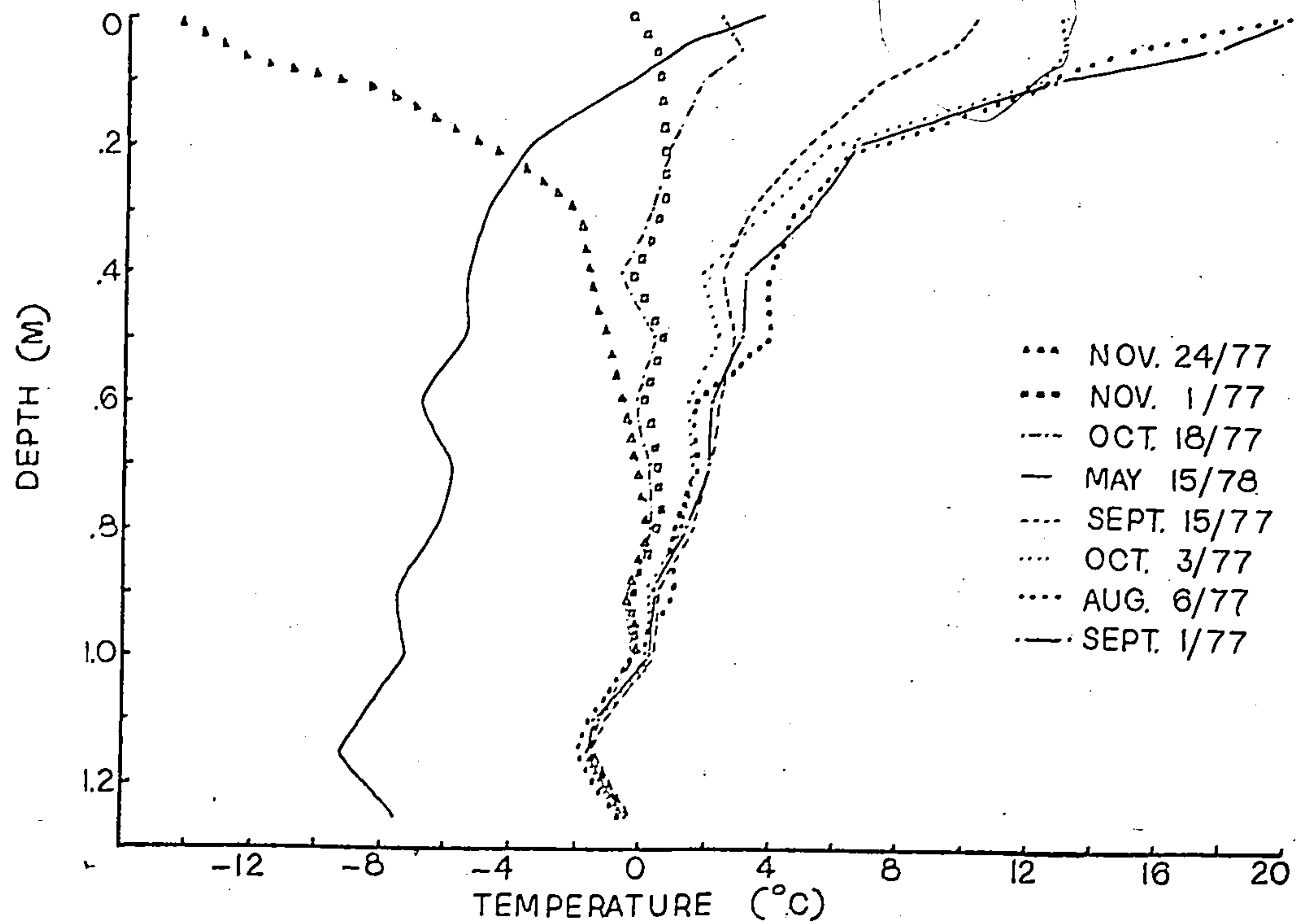


FIG. 9 SOIL TEMPERATURE PROFILES, SITE A, 1978

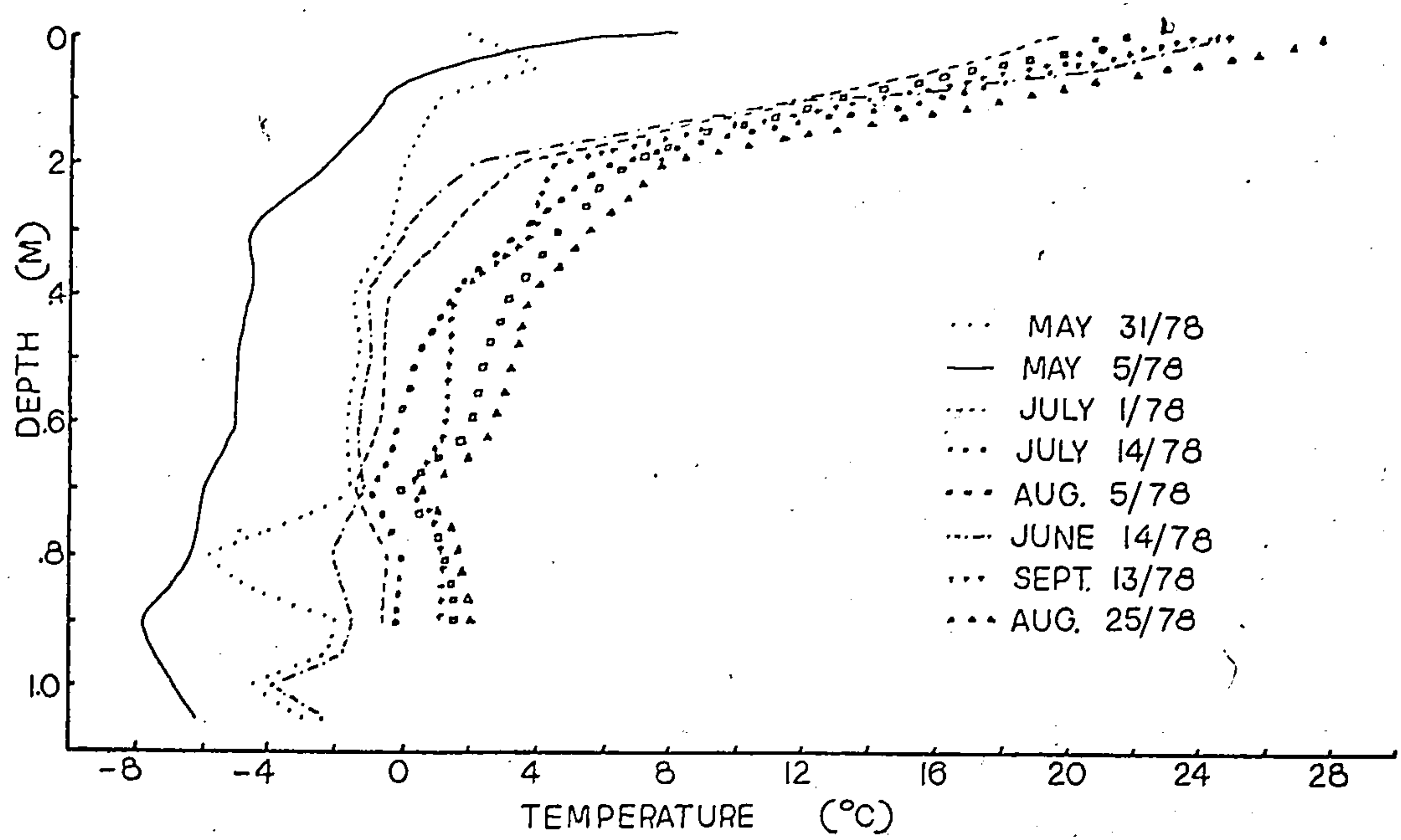
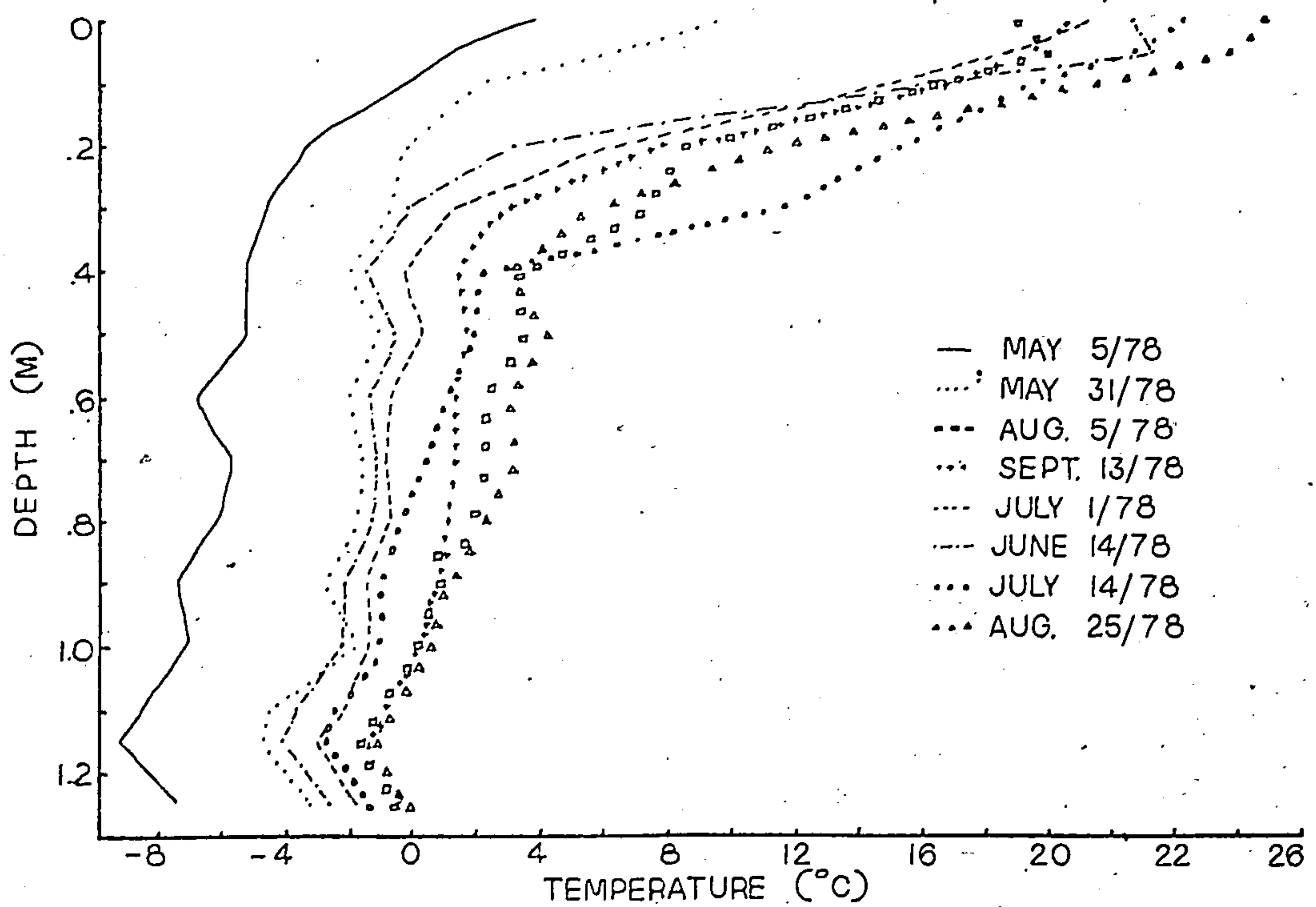


FIG. 10 SOIL TEMPERATURE PROFILES, SITE B, 1978



Soil Moisture

Figures 11 and 12 show that soil moisture averaged 30 to 32% by volume during 1977 and never varied by more than 1 to 2%. Soil moisture was fairly constant with depth. Standing water was common after rainfall and water could be squeezed from thawed soil at almost all times throughout each season. The most pronounced change in soil moisture occured during snow melt when soil moisture near the surface was as high as 44%. Standing water remained until mid-June. It is assumed that small amounts of water infiltrated into the frozen soil, while the remainder evaporated or drained off the frost mounds. Hereafter, soil moisture remained fairly constant at 31 to 32%. Moisture migration to a freezing front was not observed. During the driest periods, the surface of the lichen mat was very dry, yet soil at 5 cm remained saturated, the moss and lichen effectively inhibiting moisture loss.

The fine-grained clay soils, impermeable permafrost table, and poor drainage ensure that the soils will always have more water available than could be removed through evapotranspiration.

Radiation

Daily net radiation averages are shown in Figure 13. They fluctuate with varying surface albedos and cloud cover related to 2 to 4 day intervals of poor weather. In 1977 Q* exhibits an overall decrease

FIG. 11 SOIL MOISTURE, 1977 (PERCENTAGE BY VOLUME)

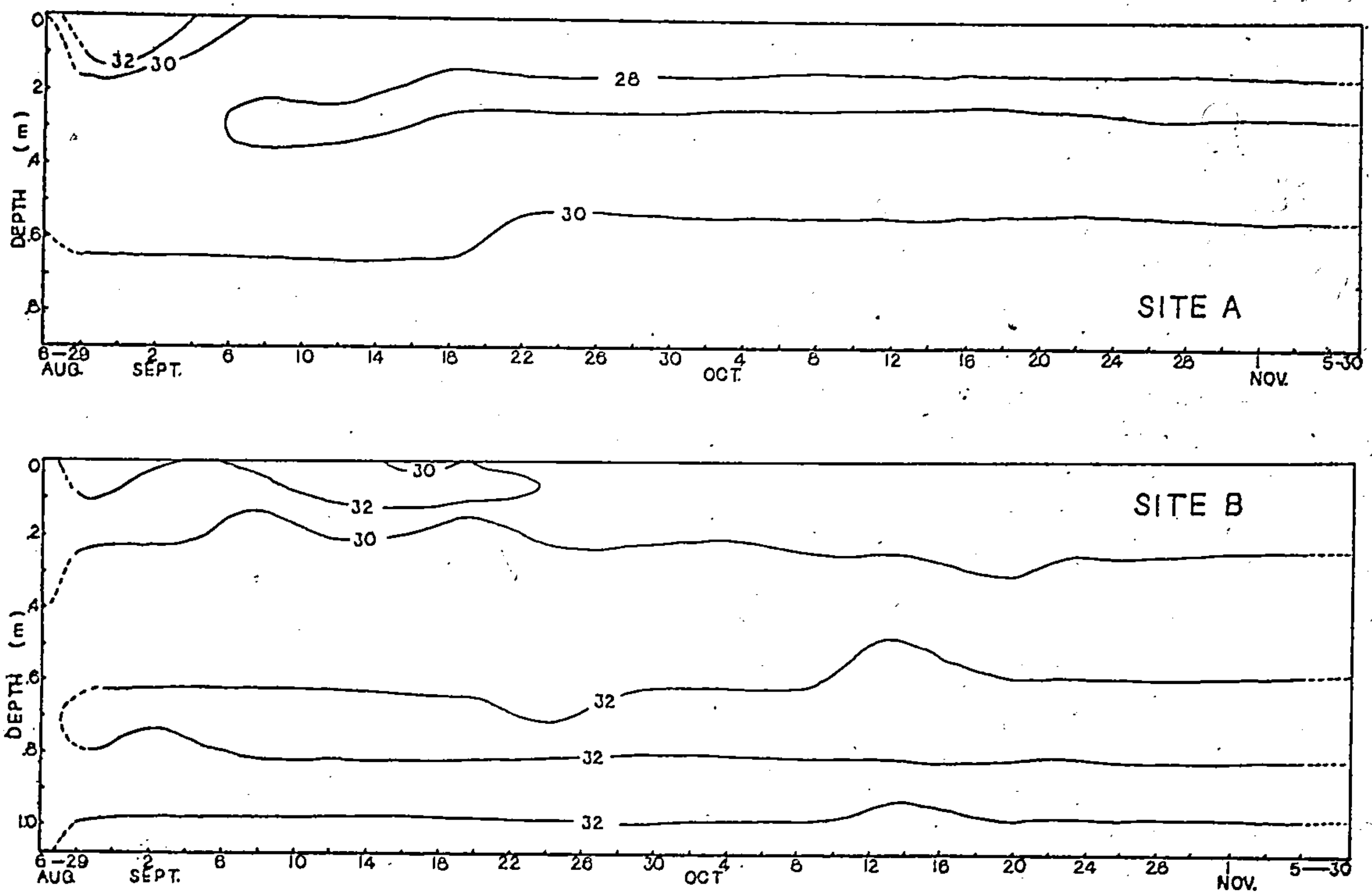


FIG. 12 SOIL MOISTURE, 1978
(PERCENTAGE BY VOLUME)

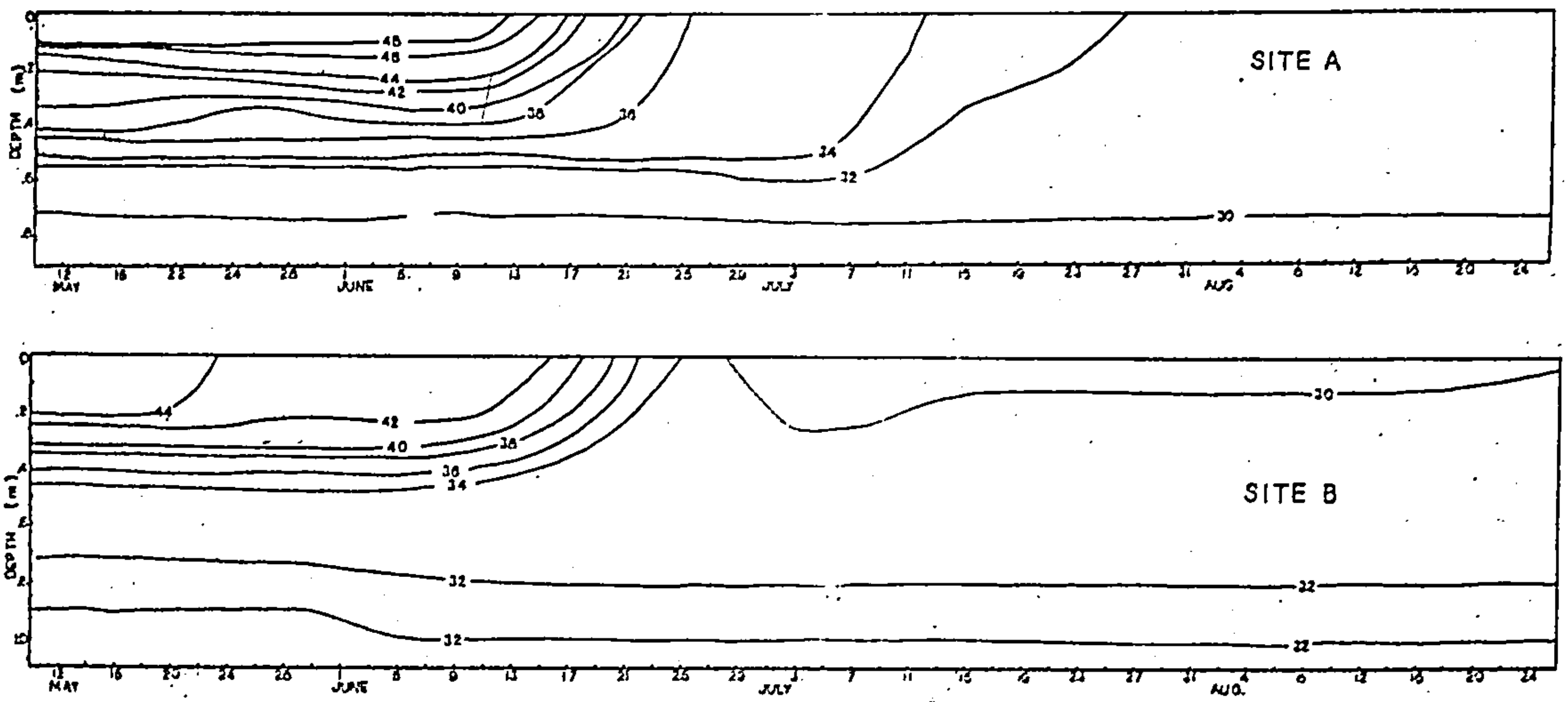
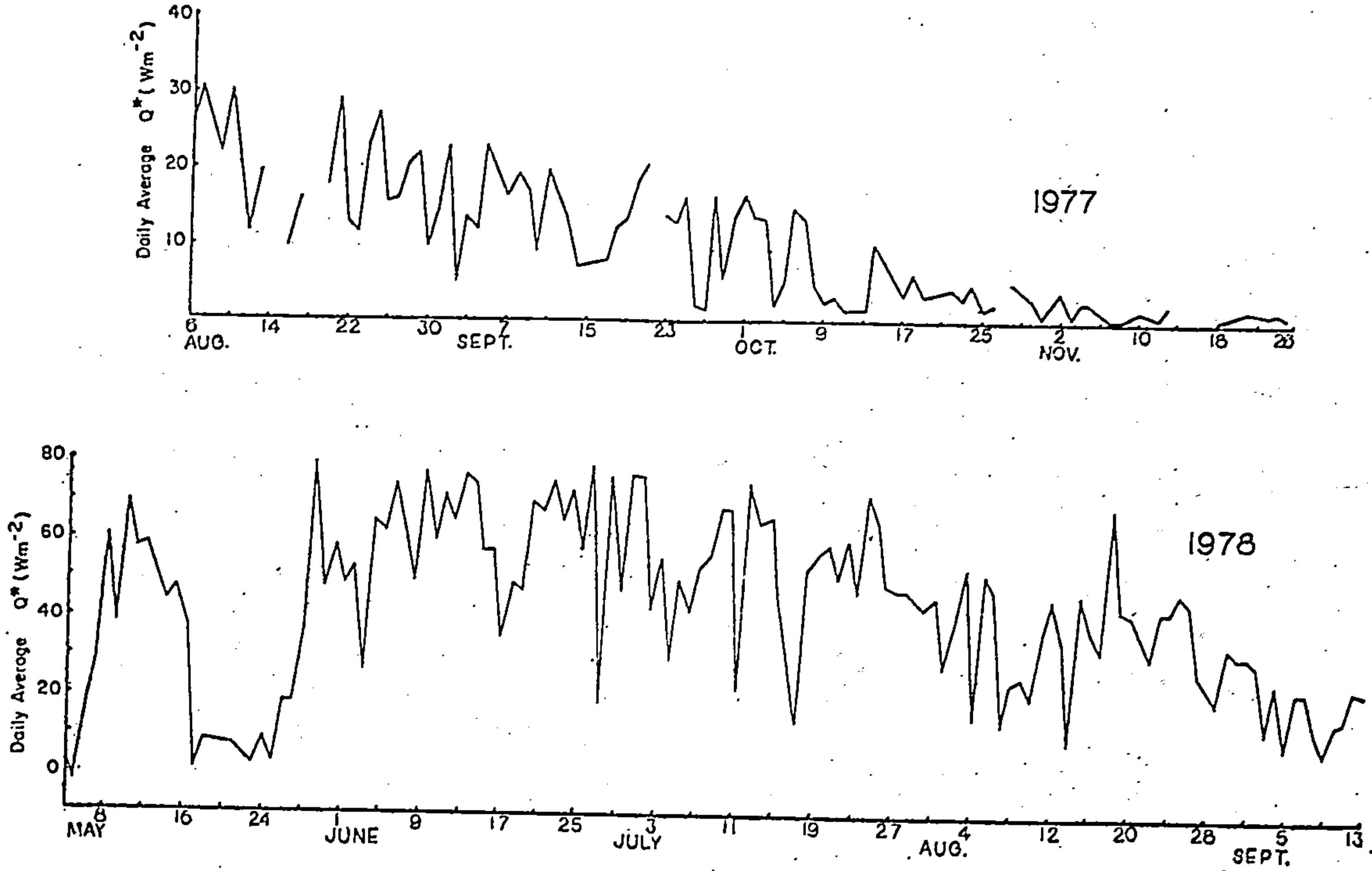


FIG. 13 DAILY NET RADIATION



with time due to shortening daylight hours. The small averages for October 11 to 13, 1977 are due to the high albedo imposed by a fresh snowcover. Snowcover is present continuously after November 13. The effect of snow cover on albedo is very apparent in May 1978 with the longer daylight hours. May 5 to 17 was a period of steady snowmelt with conditions ranging from patchy to completely bare, giving average daily net radiation values of 40 to 70 Wm^{-2}. A heavy snowfall on May 17 restored a continuous snowcover, increasing the albedo from 0.2 to 0.9, and dramatically decreasing daily net radiation. As the snowpack ripened and melted Q* gradually increased. Weller and Holmgren (1976) found that Q* at Barrow, Alaska was most influenced by the high albedo of seasonal snow cover, and low summer stratus cloud. Combinations of these factors could cause Q* to fluctuate between 10 and 85% of potential. Such fluctuations in Q* at the Churchill study sites are evident in Figure 13. The highest daily Q* values are centered about the June 21 summer solstice when daylight lasts 18 hours. Daily averages decrease steadily thereafter.

Frost Heave

Of the 6 frost heave tubes installed, 4 suffered human disturbance in the fall of 1977. As of November 24, 1977 no heave was observed, but by May 3, 1978 the remaining 2 tubes showed 4.3 and 4.4 cm of heave.

If the average moisture content of the 100 cm thick active layer was 32%, the volume expansion of water on freezing, a factor of 1.09, would result in a frost heave of 3 cm. Since the measured frost heave is greater than this it is assumed there was some moisture migration to the freezing front.

Soil Heat Exchange

The examination of the heat exchange between soil and atmosphere emphasizes the role of water. The great quantity of latent heat involved in the freezing or thawing of saturated soils plays a key role in seasonal energy exchanges. While heat exchange from soil solids depends on the temperature change and heat capacity, the exchange from soil water depends on the volume of water and its phase changes. The latent heat released upon freezing of soil water counterbalances heat loss from the soil and can even cause slight warming. This additional heat must be removed to allow for soil cooling below the freezing point.

The exchange of latent heat can be very large compared to the other soil heat exchanges. For example, G_L can exceed Gs + Gw by an order of magnitude in mid-summer (Figures 14 and 15). Latent heat exchange accounts for 84% (Site A) and 80% (Site B) of the total soil heat exchanged in both study periods. If the study had continued through the winter such that the profiles froze completely and experienced prolonged winter cooling, these values would alter. Lewis and Callaghan (1976) found that latent heat accounted for 95 to 98% of total exchange.

FIG.14 COMPONENTS OF SOIL HEAT EXCHANGE, SITE A

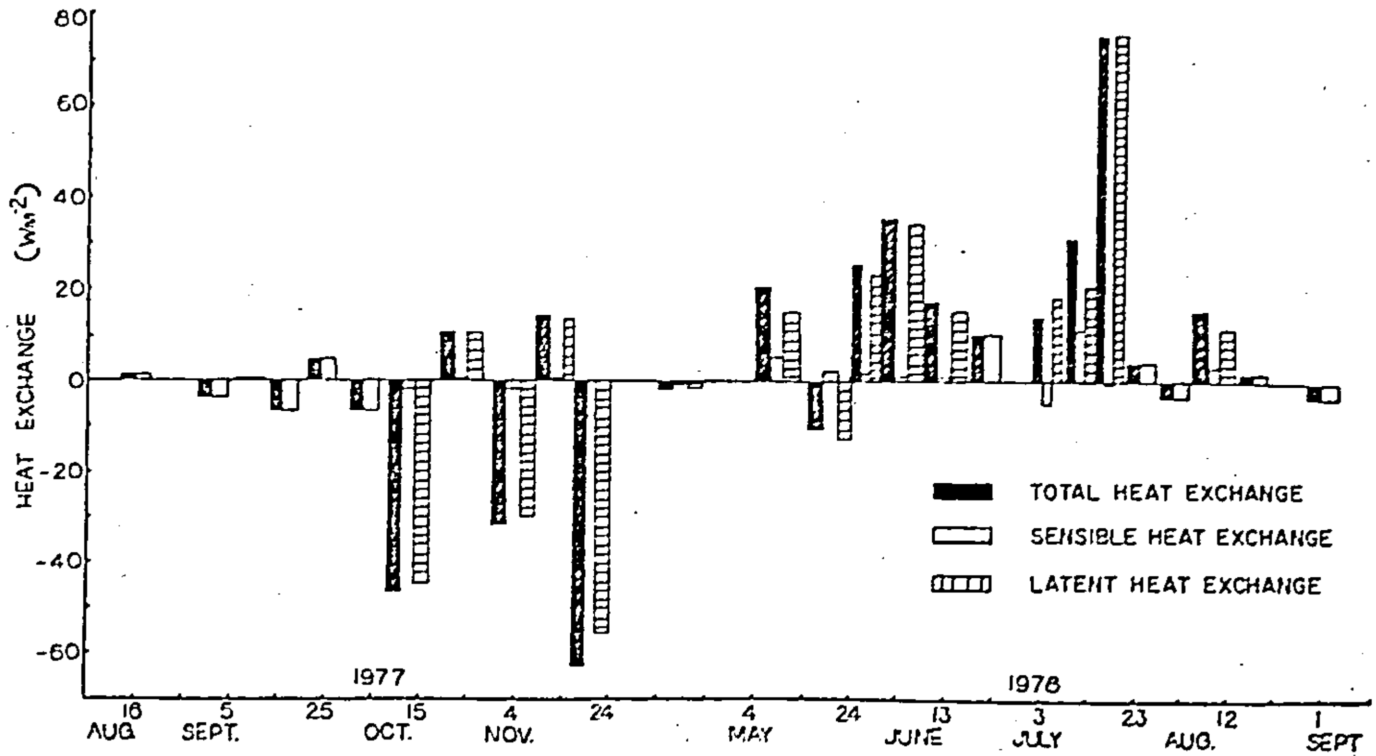
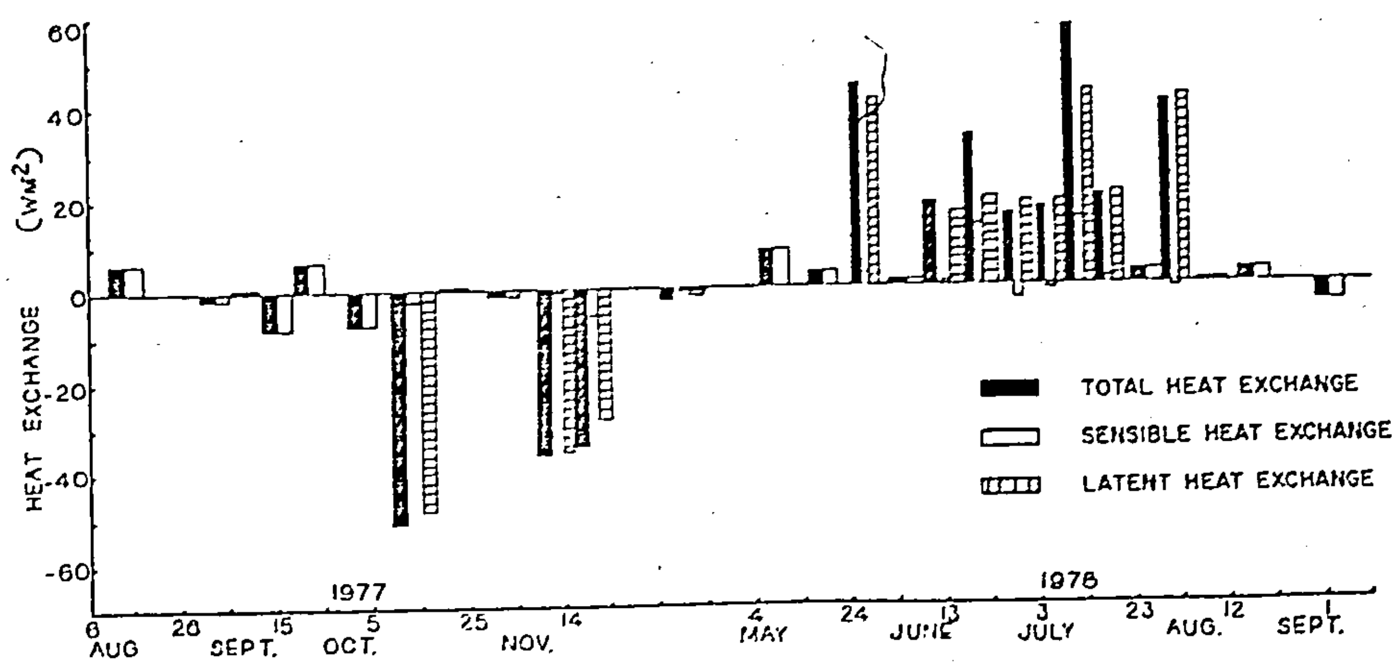


FIG. 15 COMPONENTS OF SOIL HEAT EXCHANGE, SITE B



Soil heat exchange from soil water and solids $(G_s + G_w)/Q^*$ as a percentage of net radiation is shown in Table 8. The average $(G_s + G_w)/Q^*$ of 4% during snowmelt is close to the 9% measured at Barrow by Weller & Holmgren (1976). The proportion decreases to 3.5% in mid-summer. If latent heat exchange is included in these calculations, the values can be as high as 258% during freeze-back.

Figures 14, 15 and 16 and Tables 9 and 10 show that the periods of most active heat exchange occur during freeze-back and thaw when latent heat exchanges occur. The deeper and wetter the active layer, the longer the high rate of heat exchange will continue. Differences in the average daily heat exchange between sites is the result of various layers freezing or thawing during different time periods. The greatest rate of heat exchange is in the 0 to 25 cm layer where the greatest temperature fluctuations occur. The 45 to 105 cm layer exchanges substantial amounts of heat only when freezing or thawing.

During May, 1978, the increased soil moisture due to infiltration during snow melt (Fig. 12) added a large amount of heat to the upper soil layers, enhancing soil warming. If the melt-water refroze in the soil, releasing latent heat, this would cause a further increase in soil temperatures.

FIG. 16 AVERAGE DAILY SOIL HEAT EXCHANGE TO 1.05 M.

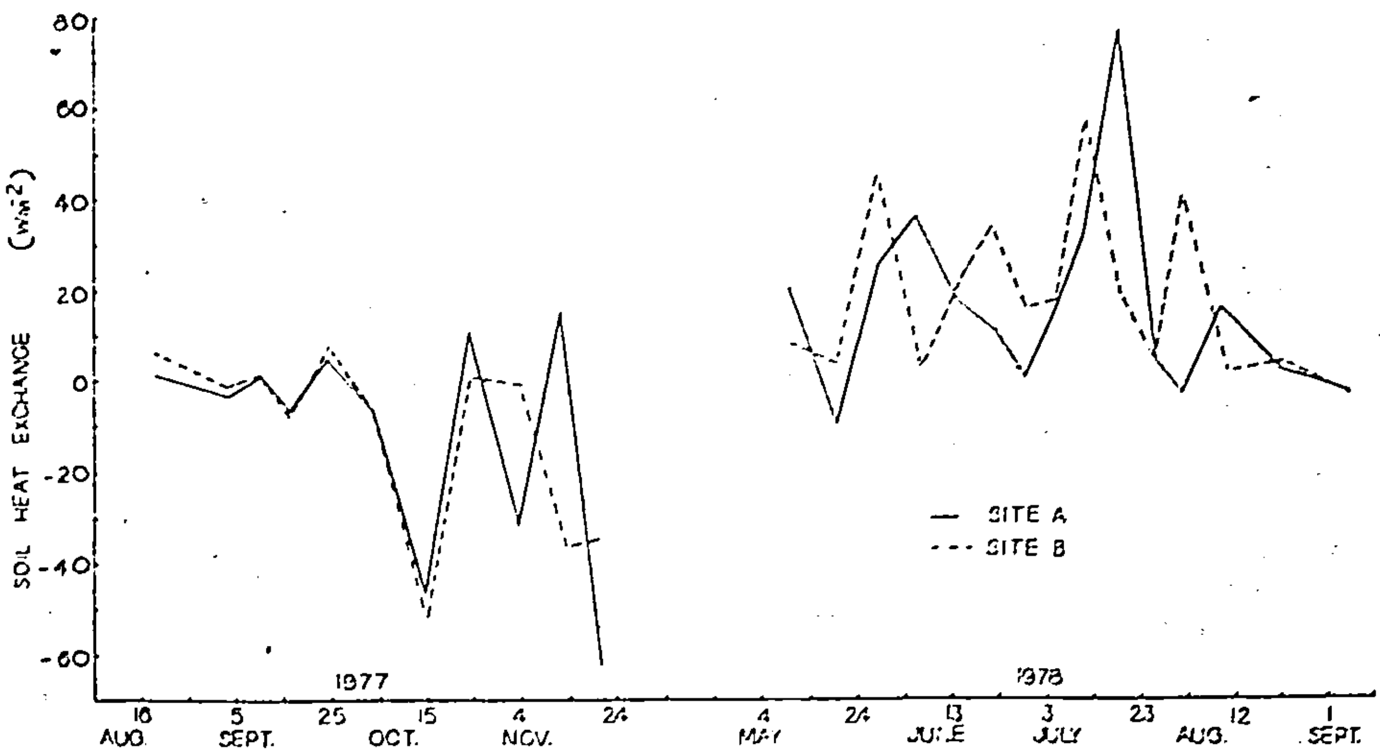


Table 8
Soil Heat (Gs + Gw) Exchange
As A Percentage of Net Radiation

Period	(Gs + Gw)/Q* (%) Site A	Site B
1977		
Aug. 6-29	+1.0	+1.0
Aug. 29-Sept. 6	-3.0	-1.0
Sept. 6-12	+1.0	+1.0
Sept. 12-19	-8.0	-10.0
Sept. 19-28	+4.0	+6.0
Sept. 28-Oct. 8	-8.0	-8.0
Oct. 8-18	-5.0	-6.0
Oct. 18-29	-0.3	+1.0
Oct. 29-Nov. 8	-7.0	-4.0
Nov. 8-16	+1.0	-2.0
Nov. 16-24	-30.0	-24.0
1978		
May 4-13	+5.0	+8.0
May 13-24	+5.0	+6.0
May 24-30	+2.0	+3.0
May 30-June 8	+1.0	+1.0
June 8-17	+1.0	+1.0
June 17-24	+6.0	+8.0
June 24-July 1	+0.1	-2.0
July 1-8	-4.0	-1.0
July 8-14	+7.0	+9.0
July 14-20	-0.4	-1.0
July 20-28	+3.0	+2.0
July 28-Aug. 3	-3.0	-1.0
Aug. 3-13	+4.0	+1.0
Aug. 13-25	+2.0	+3.0
Aug. 25-Sept. 13	-5.0	-6.0

Table 9
Components of Soil Heat
Exchange, Site A
(to 1.05 m)

Period	Average Daily Rate (Wm^{-2})			
	G_W	G_S	G_L	$G_{(total)}$
1977				
Aug. 6-29	+0.72	+0.69		+1.41
Aug. 29-Sept. 6	-1.98	-1.81		-3.79
Sept. 6-12	+0.53	+0.25		+0.78
Sept. 12-19	-3.34	-3.21		-6.55
Sept. 19-28	+2.61	+2.19		+4.80
Sept. 28-Oct. 8	-3.37	-3.22		-6.59
Oct. 8-18	-0.98	-1.02	-44.91	-46.91
Oct. 18-29	-0.06	+0.07	+10.89	+10.90
Oct. 29-Nov. 8	-0.92	-0.97	-29.94	-31.83
Nov. 8-16	+0.11	+0.10	+14.49	+14.70
Nov. 16-24	-2.40	-4.82	-55.54	-62.76
Nov. 24-May 4	-0.12	-0.19	-1.44	-1.75
1978				
May 4-13	+2.00	+3.25	+15.45	+20.70
May 13-24	+1.02	+1.41	-12.64	-10.21
May 24-30	+0.96	+1.03	+23.18	+25.17
May 30-June 8	+0.57	+0.65	+34.77	+35.99
June 8-17	+0.52	+0.48	+16.74	+17.74
June 17-24	+5.81	+5.10		+10.91
June 24-July 1	+0.53	-0.22		+0.31
July 1-8	-2.55	-2.01	+18.76	+14.21
July 8-14	+5.92	+5.31	+20.60	+31.83
July 14-20	-0.34	-0.16	+76.59	+76.09
July 20-28	+2.10	+1.92		+4.02
July 28-Aug. 3	-1.78	-1.67		-3.45
Aug. 3-13	+1.64	+1.58		+15.20
Aug. 13-25	+1.06	+0.93	+11.98	+1.99
Aug. 25-Sept. 13	-1.73	-1.53		-3.26

Table 10
Components of Soil Heat Exchange, Site B
(to 1.05 m)

Period	Average Daily Rate (Wm^{-2}) G_W	G_S	G_L	$G_{(total)}$
1977				
Aug. 6-29	+0.43	+0.44	+5.21	+6.08
Aug. 29-Sept. 6	-0.85	-0.79		-1.64
Sept. 6-12	+0.53	+0.45		+0.98
Sept. 12-19	-4.36	-3.94		-8.30
Sept. 19-28	+3.20	+3.04		+6.24
Sept. 28-Oct. 8	-3.71	-3.52		-7.23
Oct. 8-18	-1.37	-1.23	-48.68	-51.28
Oct. 18-29	+0.17	+0.17		+0.34
Oct. 29-Nov. 8	-0.59	-0.60		-1.19
Nov. 8-16	-0.16	-0.33	-35.73	-36.22
Nov. 16-24	-1.88	-4.02	-28.98	-34.88
Nov. 24-May 4	-0.27	-0.47	-1.59	-2.33
1978				
May 4-13	+3.21	+5.12		+8.33
May 13-24	+1.12	+2.01		+3.13
May 24-30	+1.22	+1.64	+41.53	+44.39
May 30-June 8	+0.85	+0.92		+1.77
June 8-17	+0.31	+0.60	+17.60	+18.51
June 17-24	+6.89	+6.57	+19.87	+33.33
June 24-July 1	-1.61	-1.34	+18.76	+15.81
July 1-8	-0.40	-0.44	+18.21	+17.37
July 8-14	+7.34	+7.09	+42.50	+56.93
July 14-20	-0.75	-0.68	+20.60	+19.17
July 20-28	+1.52	+1.57		+3.09
July 28-Aug. 3	-0.41	-0.57	+41.20	+40.22
Aug. 3-13	+0.50	+0.70		+1.20
Aug. 13-25	+1.71	+1.58		+3.29
Aug. 25-Sept. 13	-2.09	-1.86		-3.95

CHAPTER VI

CONCLUSIONS

This study demonstrates that soil moisture is the most significant factor affecting soil heat exchange and temperature change. The higher the moisture content, the slower the rate of thaw or freezing. The vapourization of water and melting snow at the surface require large amounts of energy which can detract from soil warming during thaw. Early winter snow cover can delay soil cooling, although snow cover at the study sites was not sufficiently deep to have an obvious effect.

The latent heat of fusion involved in water phase changes is the largest component of soil heat exchange, and the latent heat released during freeze-back delayed soil freezing at the study sites by 4 to 6 weeks. That the zero curtain effect at the study sites is longer than that reported in other Arctic locations is due to the saturated state of the soils and to a deeper active layer. The clay soils remain saturated due to poor drainage, an impermeable permafrost table and vegetation that is strongly resistant to moisture evaporation.

It is the changes in thermal conductivity, particularly of the organic layer and vegetation mat, that are responsible for maintaining the particular thermal regime. The thermal conductivity of frozen, saturated organic material, vegetation, or mineral soil is much higher than in the thawed state, and as such facilitates winter cooling. During summer the organic and vegetation layers effectively insulate the subsoil from pronounced heating, creating very steep near-surface temperature gradients.

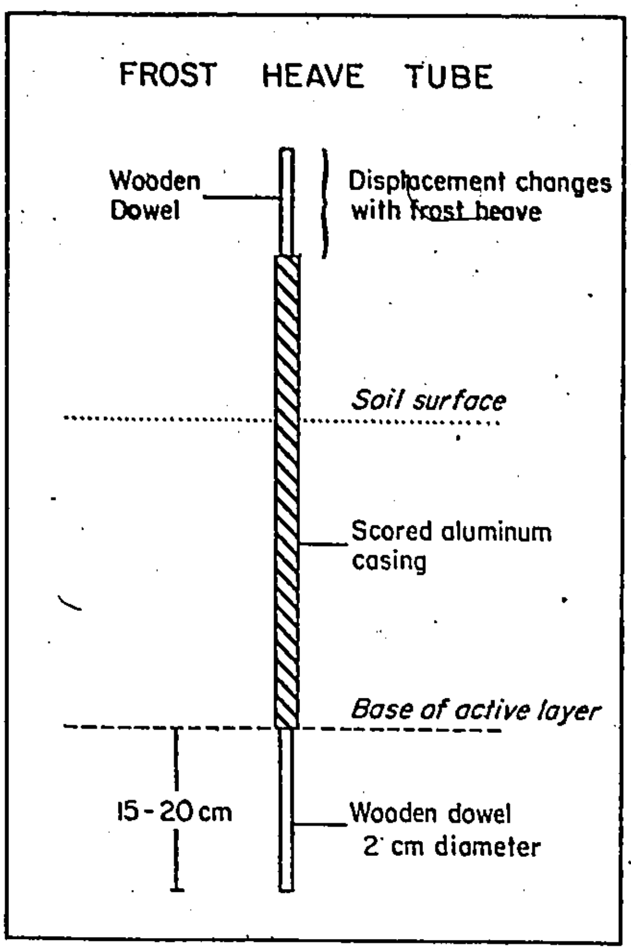
Any improvement in drainage of these soils would strongly alter the thermal regime. The heat capacity and thermal conductivity of the soil would be reduced. The rate of soil heat exchange would be affected and latent heat would contribute much less to total heat exchange, thereby shortening the zero curtain effect.

Freezing was observed to proceed both from the surface downwards and from the permafrost table up. The amplitude of temperature change with depth decreased during freeze-back, but was fairly constant during thaw, reflecting the greater conductivity of a frozen soil. Measured frost heave was close to calculated frost heave.

The amount of radiant energy at the study sites was most influenced by surface albedo. High snow albedos coupled with short daylight periods resulted in negative daily net radiation averages from early to late winter.

The thermal regime and heat exchange of wet tundra clay soils and stability of the permafrost are most sensitive to changes in thickness and quality of the organic mat and vegetation layer, and to changes in drainage conditions.

Appendix 1


FROST HEAVE TUBE
Wooden Dowel
Displacement changes with frost heave
Soil surface
Scored aluminum casing
Base of active layer
15-20 cm
Wooden dowel 2 cm diameter

REFERENCES

Balobaev, V.T., 1964: Heat exchange between permafrost and the atmosphere in the presence of a vegetation cover. NRC technical translation 1670 (1973).

Brown, R.J.E., 1969: Some observations on the influence of climatic and terrain factors on permafrost at Norman Wells, N.W.T., Can., Can. J. Earth Sci. 2: 15-31.

__________, 1973: Permafrost distribution and relation to environmental factors in the Hudson Bay Lowlands. DBR research paper 576. Proceedings of the Symposium on the Physical Environment of the Hudson Bay Lowlands, University of Guelph, Guelph, Ontario.

__________, 1978: Influence of climate and terrain on ground temperatures in the continuous permafrost zone of Northern Manitoba and Keewatin District, Canada. Proceedings of Third International Conference on Permafrost, Edmonton.

__________, & T.L. Péwé, 1973: Distribution of permafrost in North America and it's relationship to the environment: a review 1963-73, repr. North American Contribution to Second International Permafrost Conference, 1973, NRC publication 13874.

__________, & G.P. Williams, 1972: The freezing of peatland. NRC publication 12881, DBR technical paper 381.

Brown Beckel, D.K., 1957: Studies on the seasonal changes in the temperature gradient of the active layer of soil at Ft. Churchill, Manitoba, Arctic 10: 151-183.

Brady, N.C., 1974: The Nature and Property of Soils. McMillan Publishing Company, New York, pp. 639.

Drew, J.V. et al., 1958: Rate and depth of thaw in Arctic soils. Trans. of the AGU. 39: 697-701.

Geiger, R., 1965: The Climate Near the Ground. Harvard University Press, pp. 611.

Gold, L.W. and A.H. Lachenbruch, 1973: Thermal conditions in permafrost- A review of North American literature. North American Contribution to Second Annual Permafrost Conference, Yakutsk, Siberia, pp. 3-25.

Goodrich, L.E., 1978: Some results of a numerical study of ground thermal regimes. Proceedings of Third International Conference on Permafrost, Edmonton.

Jumikis, A.R., 1966: Thermal Soil Mechanics. Rutgers University Press, pp. 267.

Kelley, J.J. and D.F. Weaver, 1969: Physical processes at the surface of the Arctic tundra. Arctic 22: 426-437.

Kudryavtsev, V.A., 1959: Temperature, thickness and discontinuity of permafrost in Principles of Geocryology Part I, pp. 219-273. NRC technical translation 1187 (1965).

List, R., 1958: Smithsonian Meteorological Tables. Smithsonian Institute, Washington, pp. 527.

Martynov, G.A., 1959: Heat and moisture transfer in freezing and thawing soils in Principles of Geocryology Part I, pp. 153-192. NRC technical translation 1065 (1963).

McKay, D.C. & G.W. Thurtell, 1978: Measurements of energy fluxes involved in the energy budget of a snow cover. J. Appl. Met. 17: 339-349.

Ng, E. & P.C. Miller, 1977: Validation of a model of the effects of tundra vegetation on soil temperatures. Arct. Alp. Res. 9: 89-104.

Outcalt, S.E. et al., 1975: Computer simulation of the snowmelt and soil thermal regime at Barrow, Alaska. Water Res. Res. 11: 709-715.

Penner, E., 1972: Influence of frost rate on heaving. NRC publication 12868, repr. from Highway Research Record #393, 1972: 56-54.

Richard, W. and J. Brown, 1972: The performance of a frost tube for the determination of soil freezing and thawing depths. Soil Sc. 113: 149-154.

Rouse, W.R., 1973: The microclimatology of different terrain units in the Hudson Bay Lowlands. Proceedings from Symposium on the Physical Environments of the Hudson Bay Lowlands, University of Guelph, Guelph, Ontario.

_____________, 1976: Microclimatic changes accompanying burning in subarctic lichen woodland. Arct. Alp. Res. 8: 357-376.

_____________, & K.A.Kershaw, 1971: The effects of burning on the heat and water regimes of lichen-dominated subarctic surfaces. Arct. Alp. Res. 3: 291-304.

_____________, 1973: Studies on lichen dominated systems. VI. Interrelations of vegetation and soil moisture in the Hudson Bay Lowlands. Can. J. Bot. 51: 1309-1316.

Scott, R.F., 1964: Heat exchange at the ground surface. CRREL publication, part II: Physical Sciences, Section A: Geophysics monograph 11-Al, Hanover, New Hampshire.

Sellers, W.D., 1965: Physical Climatology. University of Chicago Press, Chicago.

Smith, M.W., 1975: Numerical simulation of microclimatic and active layer regimes in a high Arctic environment. Arctic Land Use Research Program. Dept. of Indian and Northern Affairs, Ottawa.

Stewart, R.B. and W.R. Rouse, 1976: Simple models for calculating evaporation from dry and wet tundra surfaces. Arct. Alp. Res. 8: 263-274.

Thompson, H.A., 1968: The climate of Hudson Bay in Science and History of Hudson Bay, v. 1., ed. C.S Beals, Dept. of Energy Mines and Resources, Ottawa.

Tsytovich, N.A., 1975: Mechanics of Frozen Ground. McGraw-Hill.

Tyrtikov, A.P., 1959: Perenially frozen ground and vegetation in Principles of Geocryology Part I, pp. 399-421, NRC technical translation 1163.

Vries, D.A. de, 1963: Thermal properties of soil. In Physics of Plant Environments, pp. 210-35, N. Holland Publishing Co., Amsterdam.

Weller, G.W. & B. Holmgren, 1976: The microclimates of the Arctic tundra. J. Appl. Met. 13: 854-862.

___________ et al., 1972: The tundra microclimate during snowmelt at Barrow, Alaska. Arctic 25(4): 291-300.