THE UNIVERSITY OF ALBERTA

Substrate freeze/thaw in a drained Alberta fen.

by

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Abstract

A study was undertaken to evaluate the effect of drainage on freeze/thaw cycles and frost depths in a boreal wetland site, and to determine if freeze/thaw cycles and frost depths were related to spacing of ditches or saturated hydraulic conductivity of the peat in drained areas. Literature sources suggest dry surface peat has an insulating effect. Drainage, which causes drying of surface peat, could therefore promote cooler temperatures below the ground surface and delay the thaw of ground frost.

The study was conducted in a 0.5 km^2 drained peatland located 36km southeast of Slave Lake, Alberta. The drained area contained two distinct zones, one having a ditch spacing of 25m, the other with a ditch spacing of 40m. Substrate temperature, frost thickness, frost recession and winter frost advance were sampled in the drained area and in an adjacent undrained area.

Frost was thinner in the 40m spacing area than in the 25m spacing or undrained areas. This was attributed to less water being available for freezing due to better drainage and higher saturated hydraulic conductivity of the peat in the 40m spacing area. Frost recession was faster in the undrained area than the 25m or 40m spacing areas, likely due to the higher thermal conductivity of the moist surface peat in the undrained area.

At the time of frost formation in fall, 16.7% of the sample points in the 25m spacing area contained remnant ground frost from the previous winter, versus 2.3% in the undrained area and 0% in the 40m spacing area. This was attributed to the insulating effect of the dry surface peat in the 25m spacing area. The 40m spacing area had an insulating layer of dry surface peat as well, but the frozen layer was much thinner and thus thawed more rapidly in spite of restricted heat input. Thawing occurred from both above and below the frozen layer.

Substrate temperatures were warmer at 10cm depth in the drained areas than in the undrained throughout the summer. Below 25cm, substrate temperatures were cooler in the drained areas. Diurnal temperature fluctuations at 10cm were much greater in the drained areas than in the undrained, due to the low heat capacity of the peat in the drained area.

Results of this study suggested drainage under conditions of shallow peats of low hydraulic conductivity could lower substrate temperatures enough to induce the formation of localized permafrost.

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a	Thermal diffusivity
B	Flow of heat into or out of the ground
C	Volumetric heat capacity
C_a	Heat capacity, air fraction
C_s	Heat capacity, solid fraction
C_w	Heat capacity, water fraction
k	Thermal conductivity
k_{sat}	Saturated hydraulic conductivity
L	Sensible heat exchange via mass transport
q	Heat flux into or out of the soil
S	Net radiation
t	time
T	Temperature
V	Latent heat exchange due to phase change
X_a	Volume fraction of air within the soil
X_s	Volume fraction of solids within the soil
X_w	Volume fraction of water within the soil

I. Introduction

A. Peatlands, forests, and ground frost

Canada has large areas of organic soil sites, or peatlands, located primarily in the boreal and subarctic regions (Stanek 1977). These sites are characteristic of cool, humid climates in the temperate, boreal and subarctic zones throughout the northern hemisphere (Paivanen 1984) and may also be found to a limited extent in the southern hemisphere (e.g. the highlands of Peru (Wilcox 1984)).

Many Canadian peatlands are forested; tree growth on peatlands, however, is generally poor, due to inadequate root aeration and low nutrient levels. Lowering of water tables by surface drainage is a method currently being used in the Fenno-Scandian countries in an effort to improve forest growth on peatlands. Drainage exists as a possibility for expanding the forest land base of Canada (Rennie 1977).

A possible consequence of drainage, however, could be the alteration of the thermal regime of the peat substrate. Drainage would likely cause lower moisture contents in the surface peat. As peat dries it becomes an increasingly effective thermal insulator (Brown 1963). Most of Canada's peatlands occur in areas where the substrate is at least seasonally frozen, therefore changes in freeze-thaw cycles might be expected to occur as a result of the changes in peat thermal properties induced by drainage. In light of this it was of interest to examine the relationship between peatland drainage and ground frost.

B. Literature review

1. Peatland thermal properties - historical perspectives

Relatively little attention has been paid to the study of thermal properties of peat (Burwash et-al 1971), which is surprising considering the wide distribution of peatlands in the northern hemisphere. Early studies sought to explain why substrate temperatures in peatlands were generally colder than in adjacent mineral soils. In an early Canadian study, Gagong (1897) proposed low bog temperatures were due to "a persistence of the winter cold, which in such a huge non - conducting mass would last through the summer". This represented a departure from the previously held explanation that low temperatures were due to evaporative processes unique to peatlands as a result of the large leaf surface area of the moss cover. Wilson (1939) reported peat surface temperatures closely followed air temperatures, but that at progressively lower depths temperatures dropped sharply. Rigg (1947), based on existing literature, surmised low temperatures in peatlands were due to loss of heat by evaporation at the surface as well as to the low thermal conductivity of peat.

Most of the earlier studies were done in temperate climates and were motivated by an interest in the agricultural use of organic soils. Recent research in sub-arctic and arctic climates has concentrated on the engineering problems associated with industrialization of the north. Increasing interest in oil and mineral exploration, pipelines and road construction in organic soils led to a need for a better understanding of the physical properties of organic soil sites (Macfarlane 1969). Thermal properties of these sites are of particular interest because of their importance to development and maintenance of permafrost (Brown 1968).

Permafrost, which underlies approximately one - half of the total land area of Canada, occurs extensively in peatlands (Brown 1968). Changes in permafrost depths or melting of permafrost caused by altering the overlying peat layer can have disastrous effects on construction or other projects which depend on the permafrost for structural support (Brown and Williams 1972; Macfarlane 1969). Recent work dealing with relationships between peatlands and permafrost (Brown 1963, 1968) underscored the importance of peat thermal properties.

2. Thermal properties of soils

For most soils radiant energy from the sun is the "power source" that ultimately determines the soil thermal regime. Physical factors influencing the distribution of solar radiation include latitude, exposure, climate and vegetation (Baver et al 1972). The energy available at the soil surface is the total incoming radiation minus the outward radiation flux from the soil. This quantity is known as net radiation.

The total energy exchange at a soil surface has been reduced to a "fundamental equation" by Geiger (1965). The first law of thermodynamics states that energy cannot be created or destroyed, thus at the soil - air interface the sum of all contributing factors must equal zero:

$$S + B + L + V = 0 \tag{1}$$

Where S is net radiation, B is the flow of heat into or out of the ground, L is sensible heat exchange via mass transport and V is latent heat exchange due to a phase change at the surface boundary layer. All of these factors may be positive or negative depending on whether they are inputs of heat to the surface (positive) or outputs of heat from the surface (negative).

The transfer of heat into or out of a soil (B) occurs primarily by conduction, which is the transfer of thermal energy on a molecular scale (Van Wijk and De Vries 1966). Conduction of heat in solids may be mathematically represented by a linear transport equation analogous to the transport equations developed by Fick, Darcy and Ohm for vapor diffusion, liquid and electrical conduction respectively. This equation is known as Fourier's law and states that heat flux q is proportional to temperature gradient ∇T times a proportionality constant k (Hillel 1982):

$$q = -k\nabla T \tag{2}$$

where the constant k is thermal conductivity (J/m/sec/^{o}C) and is equal to the amount of heat transferred through a unit area in unit time under a unit temperature gradient ∇T (Hillel 1982).

Fourier's law is valid only for describing heat conduction under steady-state conditions. To describe non steady-state conditions, the principle of energy conservation must be invoked, i.e. the rate of change in heat content over time of a given volume of soil equals the change in heat flux with distance (Hillel 1982). This may be represented by:

$$C(dT/dt) = - \nabla \cdot q \tag{3}$$

where C is the volumetric heat capacity (units J/m^{3}/^{o}C) and is defined as the change in heat of a unit volume of soil per unit change in temperature. Soil heat capacity is determined by adding the heat capacities of the different soil constituents weighted by their volume fractions within the soil (De Vries 1966). Thus, if C_s, C_w and C_a represent the heat capacity of soil solids, water (or ice) and air respectively, and X_s, X_w and X_a their volume fractions within the soil, the total volumetric heat capacity of the soil may be represented as follows:

$$C = X_s C_s + X_w C_w + X_a C_a \tag{4}$$

The heat capacity of each soil component is equal to density times specific heat per unit mass, i.e. C_w (the heat capacity of the water component) is the density of water times the specific heat of water (Hillel 1982).

Thermal conductivity (k) and volumetric heat capacity (C) may be combined into a single variable known as thermal diffusivity (a) where:

$$a = k/C \tag{5}$$

Thermal diffusivity expresses a soil's capacity to transmit energy from one layer to the next. In most soils, thermal diffusivity increases with increasing moisture content to a maximum and then decreases with further increases in moisture content, as more heat is required to raise the temperature of the soil water (Armson 1977).

3. Thermal properties of peat

The effectiveness of peat as an insulator is well known (Rigg 1947; Brown 1963; Macfarlane 1969; Moore and Bellamy

1974). One reason for this is the low thermal conductivity of peat. Two major factors contributing to low thermal conductivity of peat are porosity and water content (Walmsley 1973), both of which have a direct relationship to thermal conductivity. Since thermal conductivity is the transfer of heat on a molecular scale, heat transfer diminishes as solid to solid contact decreases (i.e. porosity increases) (Brown 1976; Wilson 1939).

The primary factor influencing thermal conductivity in peat is not porosity, however, but rather what the pore spaces are filled with - water, air, or ice. Thermal conductivity of water is 25 times that of air, and thermal conductivity of ice is triple that of water (Lee 1978). A dry peat layer would be expected to act as a barrier to conductive heat transfer. Heat transport into the ground in this case would be limited to convective processes, which are extremely difficult to measure (Brown 1963) and require large exposed surface areas (Brown and Williams 1972). Water-filled and especially ice-filled pores are far better conductors of heat.

The heat capacity of peat is dependent almost entirely on moisture content (Brown and Williams 1972) because the specific heat of water is high relative to the specific heat of air and peat solids. Peat soils usually contain over 80% water by volume (Boelter 1966), which makes their heat capacity high relative to mineral soils.

For soils undergoing a phase change (eg. freezing or thawing) an 'effective heat capacity' may be calculated by including in equation (4) a term accounting for heat expended or released by the phase change (Martynov 1963). The increase in 'effective' heat capacity can be dramatic. For example, 330 J/cm^3 of energy is released upon freezing of water. If no phase change occurs, this same amount of heat could cause an 80 ^oC change in temperature. Effective heat capacities of frozen peats can be as much as 100 times greater than other common soil substrates (Williams 1972), due to the much higher volumetric moisture contents of peat soils.

Since thermal conductivity (K) of peat is low and heat capacity (C) is high relative to mineral soils, the thermal diffusivity of peat, K/C, must be lower for peat. Thermal diffusivity is a measure of the rate of heat penetration into the soil, which means the depth of diurnal and annual temperature waves in peat are less than in mineral soils (Armson 1977; Brown and Williams 1972).

Evaporation has long been suspected to be a major factor in the cooling of peatlands (Rigg 1947; Gagong 1897). It has a greater effect on surface energy exchange in peatlands than in mineral soils because peatlands usually have a greater supply of water at the surface to evaporate, due to the high moisture-holding capacity of mosses as well as their ability to absorb water and bring it to the surface by capillary flow (Brown 1963). The net result is that most of the incoming radiant energy at a peat surface is used in evaporation (latent heat) and not conduction and warming at depth. The resultant drying of surface peat also inhibits heat conduction into the soil by lowering the thermal conductivity ($\bar{k}$) of the surface peat (FitzGibbon 1981).

4. Peatlands and permafrost

The process leading to permafrost formation in peatlands has been summarized by Brown (1963, 1968). During summertime, peat surface layers become dry through evaporation, thus impeding warming of the underlying soil. As autumn approaches, evaporation rates decrease, and peat surface layers become moist. When the surface layer freezes, it's thermal conductivity is increased considerably. This facilitates the transport of heat from the ground to the atmosphere and subsequently contributes to the cooling of the subsurface soil. The frozen peat cools at a greater rate in winter than the dry peat warms in summer. Thus an imbalance develops which results in a peatland area having lower mean annual below-ground temperatures than a similar area without peat cover. Permafrost will develop in areas where below ground temperatures are lowered below $0^{o}C$ throughout the year (Brown 1963).

The formation of permafrost in peatlands by the above process is highly dependent on local hydrology. In areas where surface water exists or where the water table is at the ground surface, formation of permafrost is inhibited. In both of these cases, the insulating layer of dry peat is not present in the summer and thus the thermal imbalance between summer and winter conditions is diminished or eliminated.

5. Peatland drainage and forestry

Forest growth in virgin peatlands is generally poor, due primarily to poor root aeration, lack of nutrients, and low temperatures. Experience in the Fenno-Scandian countries has shown that drainage of peatlands can increase yields and expand the forest land base (Heikurainen 1982). Profitability of drainage is dependent upon site fertility, quality of growing stock at time of drainage and the macroclimate of the site (Paivanen 1984). Macroclimate, particularly temperature, has been found to be a major factor in achieving good post-drainage growth (Heikurainen 1982).

Peatland forestry is still in experimental stages in Canada (Zoltai 1983). In eastern Canada, where black spruce is a major pulpwood species, drainage shows promise as a method of expanding the forest land base and improving yield and productivity (Rennie 1977). The climate of eastern Ontario is similar to that of Finland, becoming more continental in the northern and western portions of the province (Jeglum 1985).

A pilot study by Paivanen (1980) suggested a potential for increased tree growth with drainage in Alberta peatlands. Makitalo (1985) has suggested that the rich fens which comprise most of Alberta's peatlands might be poorly suited for drainage because the sensitivity of their nutrient balance. Peatland drainage trials are currently being undertaken in the province by the Alberta Forest Service and the Canadian Forestry Service.

C. Study objectives

Any alterations to organic soil sites which promote drying of the surface peat layer should have significant effects on thermal regime. Studies of the effects of peatland drainage in Finland and Ireland have shown an increased liability toward frost and lower substrate temperatures as a result of drainage (Pessi 1958; McEntee 1977; Valmari 1982). Because of the significant differences in thermal properties between wet and dry peat, it was hypothesized that summer frost recession and winter frost advance should be significantly different between drained and undrained areas, with the thaw in the drained area being delayed due to reduced heat flux into the subsurface caused by the insulating layer of dry surficial peat.

The objective of this study was to evaluate the effect of drainage on freeze/thaw cycles and frost depths on a boreal wetland site and to determine if freeze/thaw cycles and frost depths were related to spacing of ditches or saturated hydraulic conductivity of the peat within the drained area.

II. Methods

A. Study Area

The site selected for study was a peatland fen complex located 36 km southeast of Slave Lake, Alberta, in Sec.16, Tp.71, R.2., W. 5 Mer. (55°10′ N, 114°15′ W). This area is characterized by forest cover of black spruce (Picea mariana (Mill.) B.S.P.), tamarack (Larix laricina (Du Roi) K. Koch) and bog birch (Betula glandulosa Michx.) underlain by brown mosses (eg. Brachythecium turgidum (C.J. Hartm.) Kindb., Plagiomnium ellipticum (Brid.) Kop., Drepanocladus spp., Calliergon giganteum (Schimp.) Kindb. and Bryum pseudotriquetrum (Hedw.) Gaertner) and Sphagnum warnstorfii (Harkonen 1985). Peat deposits range from 0 to 4 m in depth. The area is part of the Lesser Slave Lake Lowlands, and overlies shale and glauconitic silty shale bedrock of the marine Upper Cretaceous Lea Park formation (Vogwill 1978). The area lies south of the established limit of discontinuous permafrost in Alberta as reported by Fisheries and Environment Canada (1977) and is south of the 0°C mean annual air temperature isotherm as well. Occurrence of localized permafrost south of these limits has been reported, however (Lindsay and Odynsky 1965).

Climate is characterized by cold winters and short, cool summers. Mean January and July temperatures (1931 - 1960) were -16°C and 16.5°C respectively (Atlas of Alberta 1969). Mean annual precipitation is 460mm with approximately

55% occurring between April and August. Annual potential evaporation by Thornthwaite's method for lowlands in the Lesser Slave Lake area is approximately 500mm (Vogwill 1978).

The principal reason for choosing the site was the existence of a peatland drainage project. In the spring of 1984 a 0.5 km^2 portion of this peatland was drained by the Alberta Forest Service. Spacing of collector ditches within the drained area was determined using a synthetic hydraulic curves method (Toth 1986)[1], which determines an optimal ditch spacing based on saturated hydraulic conductivity, desired water table depth, and a soil storage factor. Prior to ditching, the area was stratified into three distinct zones of saturated hydraulic conductivity. One of the three zones was predominantly mineral soil and was not ditched. The other two were ditched as per the synthetic hydraulic curves method, resulting in a spacing of 25m for the lower hydraulic conductivity zone ($k_{sat} = 2.1 \times 10^{-6}$m/s) and 40m for the higher hydraulic conductivity zone ($k_{sat} = 4.0 \times 10^{-6}$m/s) (Toth 1986)[1].

B. Sampling - Summer

Sample points for frost recession were located in three distinct areas: the 25m spacing zone, the 40m spacing zone, and an undrained area 100 m north of the drained area. Line

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[1] Personal communication: Discussion, Jan. 1986

transects of 10 sample points were established in the undrained and 40m spacing zone. A line transect of 12 sample points was used in the 25m spacing zone. Line transects in the drained area were oriented perpendicular to collector ditches. The differing number of sample points between zones was a result of breakdown in augering equipment used to establish sample points.

Because of high variability in microtopography throughout the study area, each sample point included a hummock and an adjacent hollow (approx. 30 - 60cm away). Hummocks and hollows were 16 to 52 cm different in elevation. Sample points in the drained areas were located 8.5m from the ditch edge in both drained areas. Sample points in the undrained area were spaced at 10m intervals.

Frost recession was determined as described by FitzGibbon (1981). This method involved augering through the frozen peat in spring to determine total frost thickness, then monitoring the recession of the upper frost boundary throughout the summer by probing. Probing was done with a 5 mm diameter rod which passed easily through the peat but did not penetrate a frozen zone. To insure that what was stopping the penetration of the rod was in fact frozen peat and not another hard object such as a tree root or log, the rod was pushed into the peat three times within a small (approximately 0.01 m^2) area at each sample point. The assumption of this approach was that it was highly unlikely that a tree root or log would stop the rod at more than one point; frozen peat would stop the rod at approximately the same depth at all three points.

Sampling began May 9-12,1985. Depth to the upper frost boundary was measured by probing on a weekly basis throughout the summer season. Beginning June 15, 1985 a second transect of 18 sample points (with one hollow and one hummock at each point) was established in each of the three areas to determine if site disturbance from foot traffic affected earlier observations. This data was combined with the earlier observations. Frost recession was judged completed when repeated probings could not detect the presence of frost.

Substrate temperature profiles were measured at four hummocks and four hollows in each of the three areas. At each point, copper - constantan (type T) thermocouples were installed at depths of 10, 25, 45, and 60 cm below the ground surface. The thermocouples were spaced at the above intervals on a wooden dowel, which was then forced into a pilot hole in the frozen peat identical in diameter to the dowel. Temperatures were recorded biweekly throughout the summer using a handheld digital multimeter and a small electronic reference junction. During July and August a datalogger was used to monitor diurnal temperature fluctuations. Temperatures were recorded at hourly intervals continuously for approximately one week at one hummock in each of the three areas.

Precipitation was monitored using a Belfort Universal rain gauge placed at the northern edge of the drained area. Water table levels were obtained from a network of recording water wells maintained by the Department of Geology, University of Alberta.

C. Sampling - Winter

Sampling frost advance in fall and in winter required a method other than probing, as frost formation generally occurs from the surface downward. Frost tubes were used as they allow repeated measurements of frost advance and retreat at a single point.

The frost tubes used in this study were constructed as described by Rickard and Brown (1972). The device consisted of two principal parts: an outer tube of 1.6cm inside diameter PVC tubing, sealed at the bottom; and an inner tube of 0.95cm (inside diameter) clear acrylic tubing, sealed at both ends and filled with sand saturated with a 0.1% solution of fluorescein dye. When in solution fluorescein changes color from bright green to a pale pink upon freezing.

The outer tube was installed vertically into the ground with a sufficient length of the tube protruding above ground level to avoid being covered by snow. The inner tube was placed inside the outer tube such that the top of the inner tube was level with the ground surface. A string attached to the inner tube allowed removal of the inner tube. With the inner tube in place the outer tube was sealed with a removable cork stopper.

Frost was monitored by periodically removing the inner tube and noting where the fluorescein had changed color. The pink areas denoted the portion of the peat which was frozen. The inner tube was returned to the outer tube and monitored again at a later date.

Frost tubes were installed October 10, 1985. They were installed at or very near to the frost recession sample points, on both a hummock and a hollow. Seven sample points were used in the undrained area, and eight in each of the drained areas (25 and 40m spacing). The frost tubes were monitored on a monthly basis throughout the winter and spring of 1985-86 and into the summer of 1986.

An attempt was made to install the tubes in the spring of 1985 and use them to monitor summer frost recession. Tubes installed in spring did not work. A probable explanation for this is that the tubes needed direct contact with ground frost to work effectively. Tubes installed in spring are likely separated from the surrounding ground frost by a thin air layer as a consequence of installation. Since the frost is thawing, this air layer increases in thickness with time. Tubes installed in fall are in contact with the soil at time of freezing, and therefore freeze in solidly.

Temperature measurements for November and December 1985 were prevented by failure of the electronic reference junction on very cold days. Temperature measurements were resumed in January 1986. Keeping the reference junction in a warm pocket enabled one to use it in the final months of winter, however performance was erratic and the accuracy of temperature measurements made during this period was questionable.

III. Results and discussion

A. Hydrologic and meteorologic data

Water table height was 20 to 50 cm lower in the drained area than the undrained. Average water table levels for the summer were 55 and 20 cm below the ground surface for the drained and undrained areas respectively. Water table responses to precipitation in all areas were pronounced, with recession of the water table to prestorm levels within 4-5 days. May through October rainfall for 1985 was 207 mm, 40% lower than the 63 year average of 350 mm for the Slave Lake region. Snowpack reached a maximum average depth of 42cm in February, 1986 (Figure 1). By April 5, 1986 only a few small patches of snow remained.

B. Frost thickness.

Surveys for total frost thickness in the three areas on May 9-10, 1985 showed frost thickness in the 40 m spacing area was significantly less than in both the 25 m spacing or undrained areas (Scheffe's test, 0.05 level). Mean frost thickness was 41.6 cm in the undrained, 36.7 cm in the 25 m spacing and 22.4 in the 40 m spacing (Table 1). Frost was thinner in hollows than in hummocks in all areas. This difference was most pronounced in the 25 m spacing zone, where frost thickness in hummocks averaged 6.4 cm thicker than hollows, and least pronounced in the undrained area,

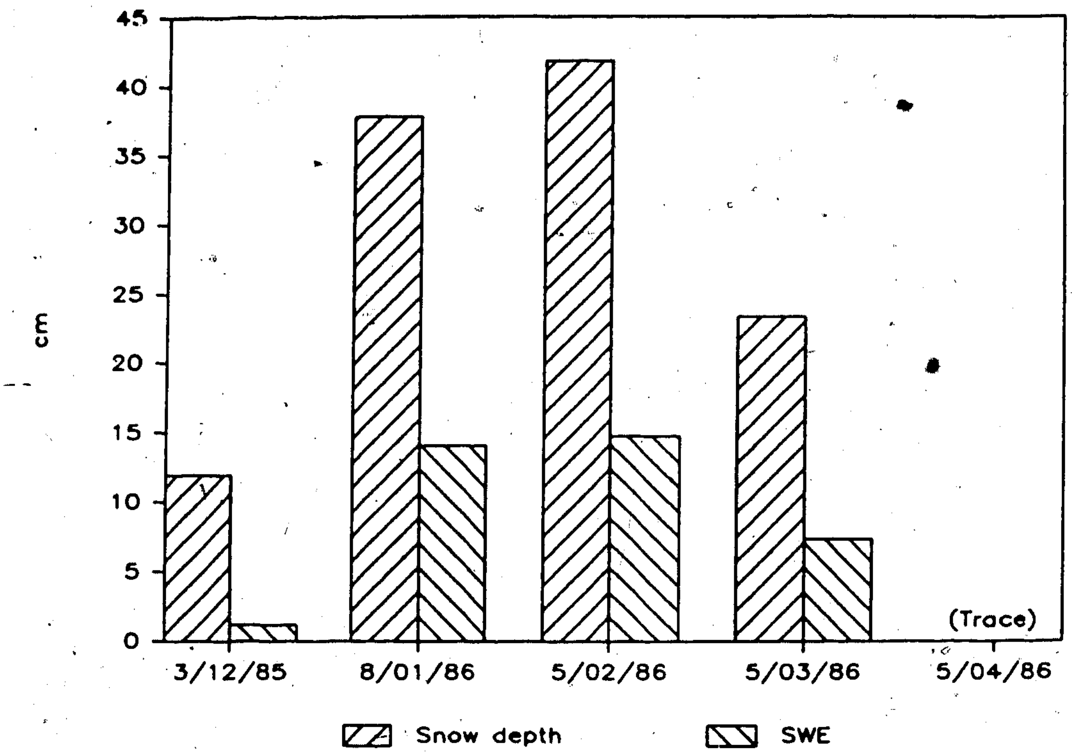

Figure 1. Snow depth and water equivalent at the study area, winter 1985 - 1986.

with a difference of 3.1 cm. The 40 m spacing zone had an average hummock/hollow frost thickness difference of 4.6 cm. It should be noted that some thawing appeared to have occurred prior to May 9, so these values may not represent the maximum winter frost thickness.

The relative ease with which the auger bit was able to bore through the frozen peat as well as the condition of the material brought up by the drill bit gave some indication of the porosity of the frozen peat in the different areas. Material from the 40 m spacing was often granular and

Group	Mean(cm)	Std. error	95% C.I.
Undrained hollows	40.0	3.2	32.8 to 47.2
Undrained hummocks	43.1	4.3	33.4 to 52.8
Dr.-40 hollows	20.1	3.2	12.9 to 27.3
Dr.-40 hummocks	24.7	3.2	17.4 to 32.0
Dr.-25 hollows	33.5	3.7	25.4 to 41.6
Dr.-25 hummocks	39.9	3.8	31.5 to 48.3
Combining hummocks and hollows:			
Undrained	41.6	2.6	36.0 to 47.0
Dr.-40	22.4	2.3	17.6 to 27.2
Dr.-25	36.7	2.7	31.2 to 42.2

Scheffe's test (0.05 level) for combined hummock and hollow means:

Dr-40	[illegible]	Undr.
22.4	36.7	41.6

Means not underscored by the same line are significantly different.

Table 1. Frost thickness of each area (May 9-10, 1985) and results of Scheffe's test for area means.

the auger passed through easily. In the 25 m spacing and the undrained zones boring was difficult and concrete ice was evident.

Frost was thinner in the 40 m spacing area than in the undrained area. This was attributed to high K_{sat} and better drainage which made less water available for freezing. Peats of high K_{sat} contain large, easily drained pores. As a result, they release more water with drainage than those of lower K_{sat} (Walmsley 1977). The 40 m spacing area, having higher K_{sat}, would be expected to retain less water with drainage than the 25 m spacing zone. Thus less water would be available for freezing in the fall, resulting in a thinner frozen layer. This suggests, however, that the spacing of ditches is of lesser importance than the physical properties of the peat. The 25m spacing area has the more intensive ditching pattern, but on the basis of frozen layer thickness the 40m spacing area appears to be better drained.

C. Frost recession

Frost recession was fastest in the undrained area. This was especially evident in the hollows (Figure 2). During the period May 10-June 6, 1985, average depth to the frost table increased 16.9 cm in the undrained area hollows, versus 7.4 cm and 7.8 cm for the 25 and 40 m spacings respectively. The corresponding averages for the hummocks were 17.5 cm, 11.8 cm and 9.8 cm.

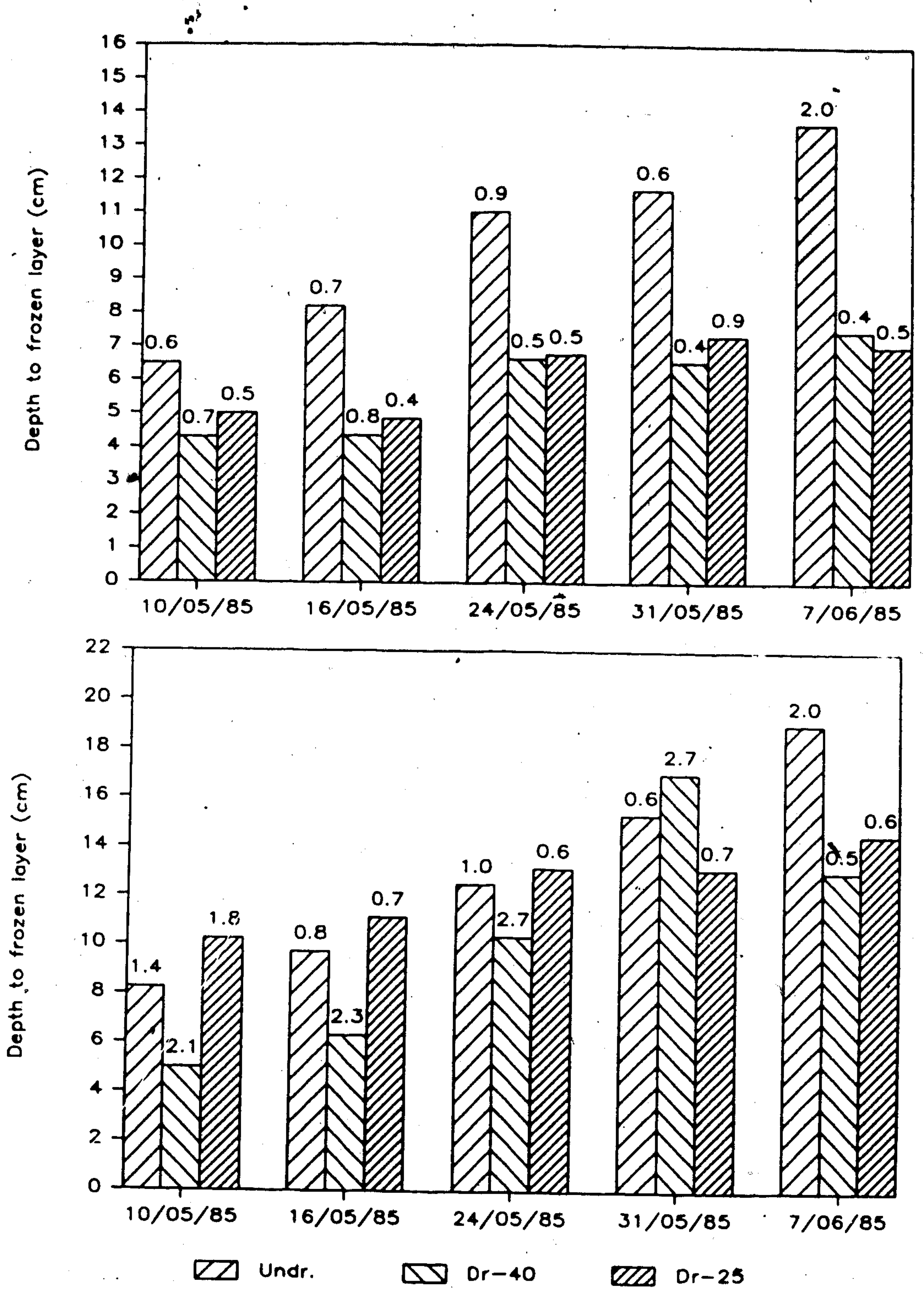

Figure 2. Seasonal depth to frost in hollows (upper graph) and hummocks (lower graph), Spring 1985. Standard errors are shown above bars.

Frost recession was slower in the drained areas than in the undrained (Figure 2) because the dry surficial peat acted as a barrier to conductive heat transport into the soil. The effectiveness of dry peat as an insulator is well documented (Rigg 1947; Benninghoff 1952; Brown 1963; MacFarland 1969; Moore and Bellamy 1974). Thermal conductivity of peat decreases greatly with reduced water content (Lee 1978), therefore heat transport into the dry surficial peat would be limited to convective processes (vapor diffusion and rainfall), which can be slow and are extremely difficult to measure (Brown 1963).

Frost recession was faster in the undrained area even though ice thickness was greater. The faster recession was attributed to the assumed higher water content and resultant higher thermal conductivity of the surface peat in the undrained area. Differences in water content of the surface peats between the drained and undrained areas may have been amplified by the unusually low rainfall observed during the study season. In the drained area the surface peat dried out as a consequence of lower water table levels, and rainfall events were too infrequent to have a significant wetting effect. Higher water table levels in the undrained area may have been sufficient to maintain a moist layer of surface peat in spite of low rainfall. It is also possible, however, that low rainfall levels could have had the opposite effect, i.e. causing drying of surface peat in all areas and thus minimizing differences between drained and undrained. Continuation of the study through several years would help clarify this.

Frozen soil was detected more frequently in the 25 m spacing zone throughout the summer than in the other areas (Figure 3). The rate at which zones became ice-free was quite similar between the undrained and 40 m spacing zones. The 25 m spacing zone retained ice much longer; indeed at the onset of frost formation in early October, 16.7% of the sample points in the 25 m spacing still retained ice. In the undrained area frozen soil was found at 2.3% of the sample points at this date (Figure 3). Frozen soil was not detected in the 40 m spacing zone after September 10, 1985. Transects started in May showed a slightly higher thaw rate than those started in June, suggesting the sites did not remain totally undisturbed during sampling.

The probable explanation for the more rapid disappearance of frost in the 40m vs. the 25m spacing area was that frost was thinner and more porous and therefore required less heat input to melt than the thicker frost layer in the 25 m spacing. The similar rates of thaw between the undrained and 40 m spacing areas are misleading - they were similar, but for different reasons. The 40 m spacing zone had a thinner frozen layer requiring less heat to thaw; the undrained area had a thicker frozen layer but received more heat input because of the higher thermal conductivity of it's surface peat. The net result was that the rates of disappearance of ice were similar between the two areas. The 25 m spacing zone had a thicker frozen layer which received limited heat input due to the low thermal conductivity of it's dry surface peat. Thus frozen conditions persisted longer than in the other areas.

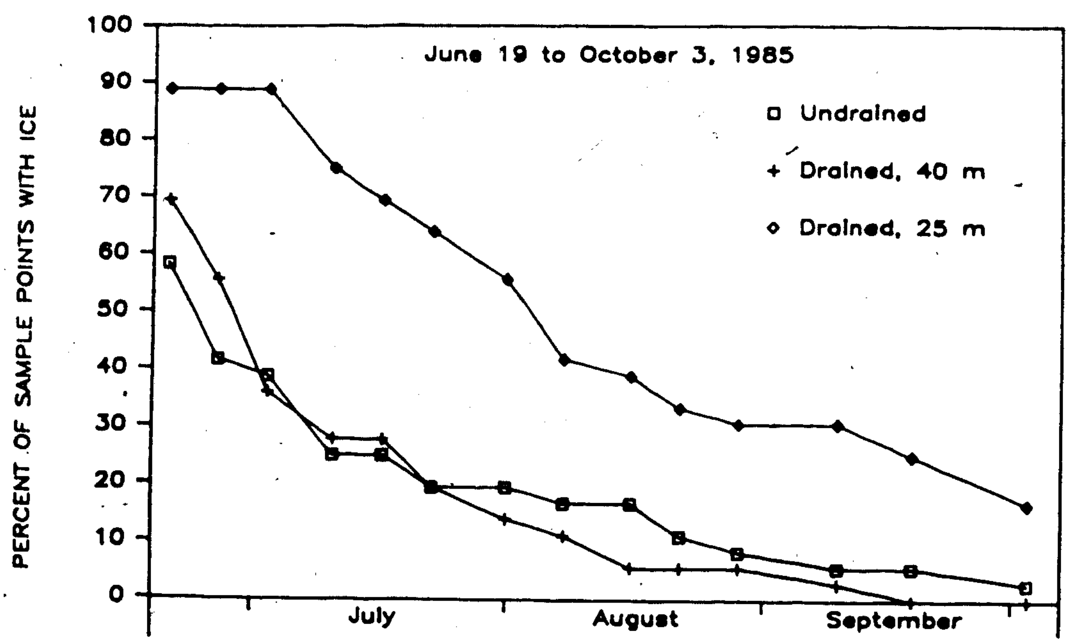

Figure 3. Percentage of sample points with ice for the period June 19 - October 3, 1985 (combined hummocks and hollows).

The presence of remnant frost in the undrained area in early October (at 2.3 percent of the sample points) was partly attributed to low summer precipitation in 1985. Summer rainfall for 1985 (May-October) was 40% below the 63 year average for the Slave Lake region. As a result, some surface peat material in the undrained area dried out, albeit to a much lesser extent than occurred in the drained area. In a wetter year frost recession would probably occur earlier and at a faster rate on all sites because of increased thermal conductivity of the surface peat as well as convective heat transfer from precipitation into the peat substrate (Ryden and Kostov, 1980).

There is some debate as to whether the sample points in the 25 m spacing zones which contained frozen peat throughout the summer, fall and following winter should be classified as permafrost. Some authors maintain that for an area to be classified as permafrost, it must remain in a frozen state for at least two years (Brown and Johnston 1964). Others (Zoltai 1986)[2] claim that a single year is sufficient. Radforth (1962) coined the term "climafrost", referring to ice that is temporary but lasts more than one year. A term that is more useful in describing the phenomena observed in this study is "localized permafrost". Localized permafrost (Zoltai 1971, 1975) refers to small isolated permafrost lenses occurring at the southern fringe of the

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[2] Personal communication: Discussion, Jan. 1986

discontinuous permafrost zone.

The localized permafrost observed in the study area is of interest because it occurs well south of established southern limits of discontinuous permafrost as well as the $0^{o}C$ mean annual air temperature isotherm (Figure 4). The formation of permafrost in peatlands is strongly associated with the insulating properties of dry peat (Brown 1963); therefore it is not surprising to find that the vast majority of the observed localized permafrost was in the drained 25 m spacing zone. This zone had a layer of dry peat as a consequence of drainage as well as a relatively thick frozen zone due to low hydraulic conductivity, and these two factors may have contributed to preservation of a frozen layer into the fall. Without data on frost occurrence prior to drainage it is impossible to tell if the localized permafrost observed in the drained-25 area is a direct result of drainage. The much greater extent of frost in this area versus the undrained, however, suggests this might be the case.

D. Thaw from beneath the frozen layer

Thaw occurred from both the top and bottom of the frozen layer in all areas, with thaw from below accounting for as much as 38% of total thaw in the undrained area. Table 2 shows the maximum depth of frost penetration as determined on May 9-10, as well as the total depth of thaw

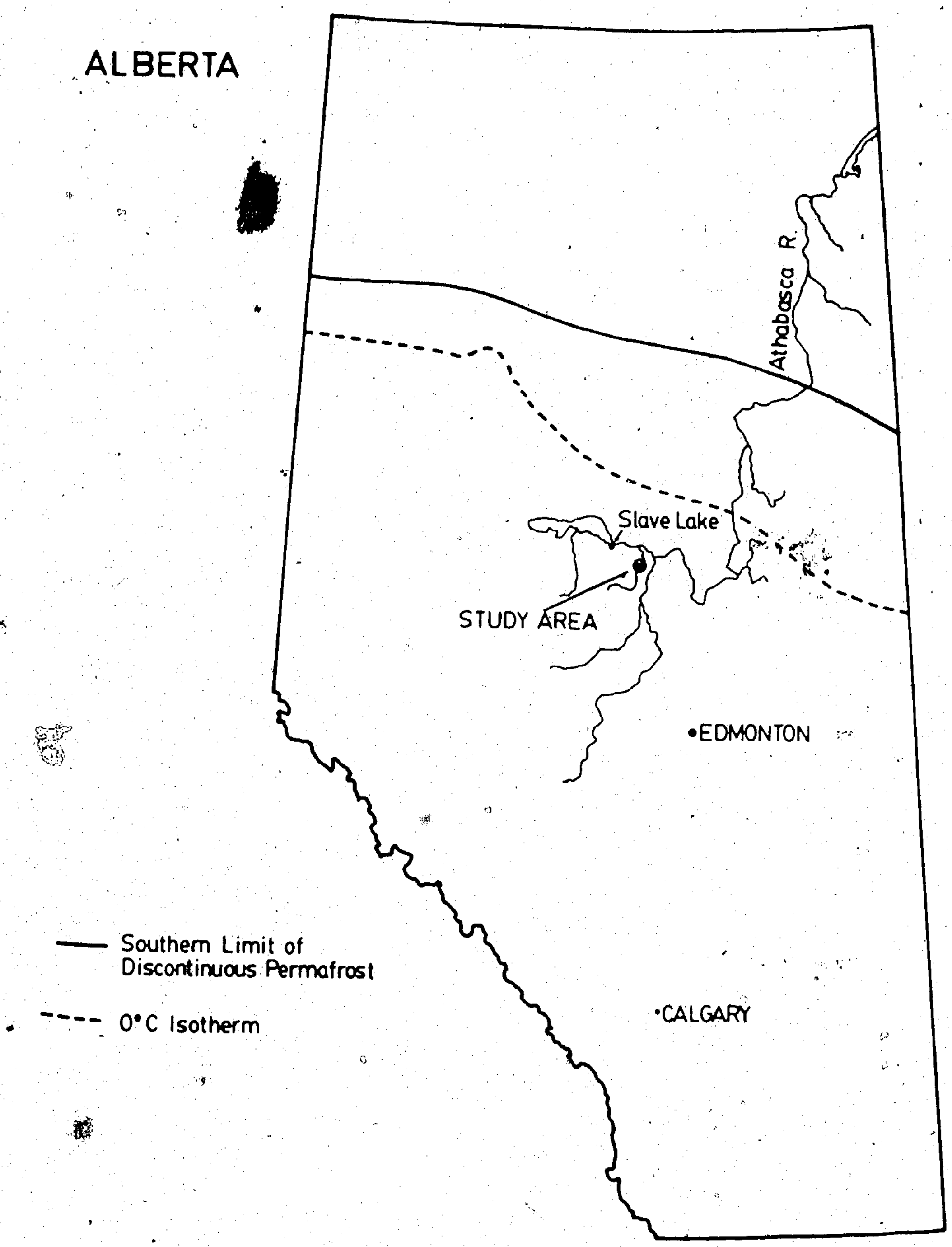

Figure 4. Study area location in relation to 0°C isotherm and southern limit of discontinuous permafrost (Fisheries and Environment Canada, 1977).

from above, and the resulting percentages of thaw which must have occurred from above and underneath. The depth of thaw from above was the position of the frozen layer one week prior to total thaw. Percent thaw from above is (thaw from surface/initial depthx100), and percent from below is (100-percent thaw from above). The final frost depth a week prior to complete thaw was indicative of the total depth of thaw from above since, in most cases, the frozen layer was so thin at this point that it could be penetrated with the probe using only light finger pressure.

The study area was located in a fen complex, therefore subsurface groundwater flow should be expected. This offers an explanation for the occurrence of thawing from underneath the frozen layer. FitzGibbon (1981) found 40% of thaw to occur at the bottom of the frozen layer in a fen in Saskatchewan, whereas in a nearby bog (with no lateral subsurface flow) no thaw from below was observed. He concluded that thaw from below was produced by heat exchange with groundwater. Thaw from below in the drained areas appeared to be less than in the undrained (27% and 31% versus 38%; see Table 2), however these differences are not statistically significant. The apparent trend of less thaw from below in the drained area might be attributed to less heat exchange with ground water due to a lower water table. Detailed data relating water table height and lower frost boundaries at specific points could clarify this.

	Initial lower frost depth (cm)		Depth of thaw from above (cm)		Percent of thaw from:	
	Mean	std. error	Mean	std. error	above	below
Undr.	56.3	(3.0)	35.0	(2.4)	62%	38%
40 m	36.9	(2.7)	26.8	(2.3)	73%	27%
25 m	52.0	(2.9)	35.9	(2.5)	69%	31%

Table 2. Percentage of thaw from above and below the frozen layer. Means are of combined hummock and hollow measurements.

E. Substrate temperature

Midday hummock peat temperatures at 10 cm depth were 0 to 4.4 °C cooler in the undrained area than in the drained area for most of the summer (Figure 5). Only in late August were midday temperatures at 10 cm depth warmer in the undrained area. Surficial peat temperatures were cooler in the undrained area because more heat energy was required to warm the peat due to it's higher water content and hence

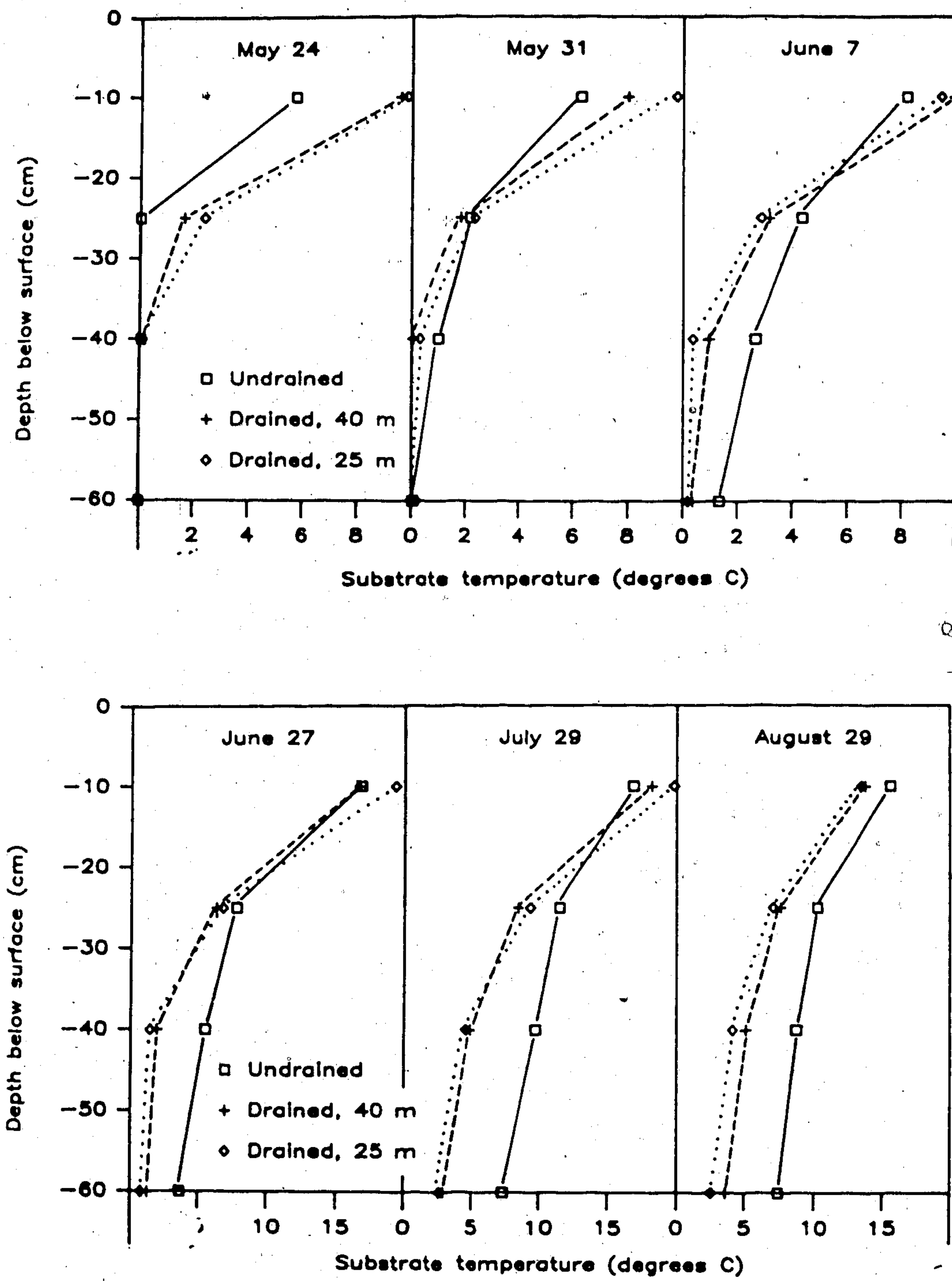

Figure 5. Midday substrate temperatures in the study area.

higher heat capacity. Higher water contents in the undrained area likely resulted in greater rates of evaporative cooling as well.

Soil temperatures below 10 cm were warmer in the undrained area than in the drained area throughout most of the summer. Differences in temperature between drained and undrained were slight at 25 cm depth. At 40 and 60 cm the differences were greater, reaching almost 5 °C at 40 cm in late July (Figure 5). Temperatures and the rate of warming at lower depths (25 - 60 cm) remained higher throughout the summer in the undrained area because of the inferred higher thermal conductivity of the moist peat.

Substrate temperature data in this study reflected the same trends as observed for frost. Temperatures were cooler in the undrained area than drained at all depths in spring due to the higher heat capacity of the moist peat in the undrained area. By early late May - early June the effect of the higher thermal conductivity of the peat in the undrained area was evident, causing warmer temperatures at 25, 40 and 60 cm depth in the undrained area than in drained.

Diurnal temperature fluctuation near the surface was greater in the drained areas than in the undrained. Fluctuations of up to 20 °C in a single day were recorded in the drained areas at the 10 cm depth (Figure 6). In the undrained area amplitudes of only 4 °C were observed at 10 cm.

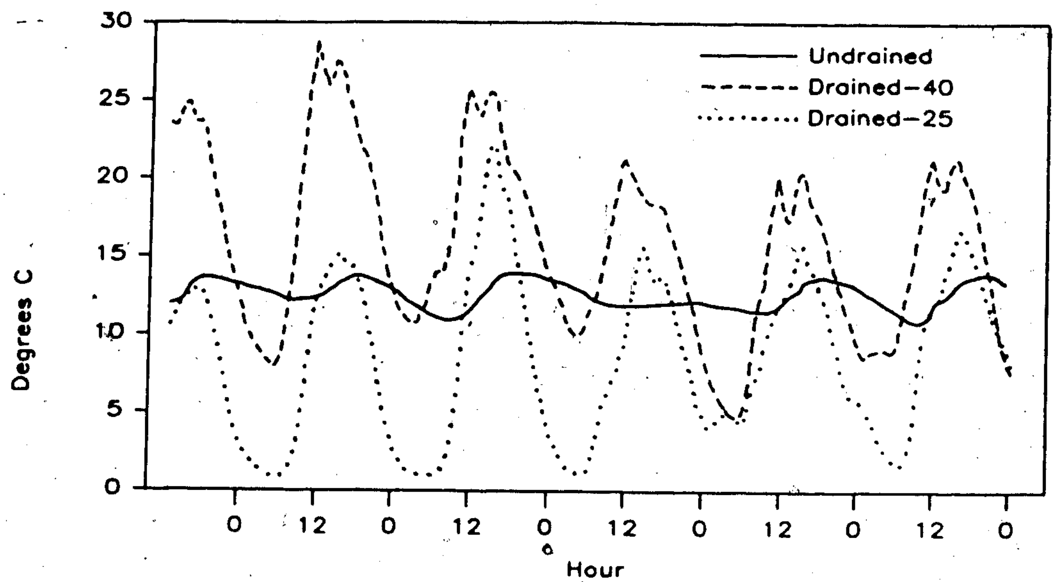

Figure 6. Diurnal temperature fluctuation at 10 cm depth for Undrained (July 23-29), Drained-40 (August 1-7) and Drained-25 (August 8-14).

At 25 cm depth diurnal fluctuations of 8 °C were recorded in the drained areas versus less than 1 °C in the undrained area. Temperature fluctuations at the 40 and 60 cm depths in all areas were small, rarely exceeding 1 °C/day.

The wide diurnal temperature fluctuation at 10 cm depth in the drained area may be attributed to the reduced heat capacity of the dry surface peat. Using equation (4), the heat capacity of the peat substrate would be equal to the sum of the heat capacities of peat materials, water and air, weighted by their volume fractions in the substrate. The specific heat of water is 333 times that of air (Hillel 1982), so as air replaces water in the substrate (i.e. as a result of drainage) the heat capacity of the peat drops dramatically.

Differences in albedo and thermal admittance among the drained and undrained areas are likely minimal due to the similarity of forest cover among the three areas. The 40 m spacing area had a slightly more open canopy than the other two areas, however, and this might have contributed to it's rapid thaw.

Differences in thermal diffusivity between drained and undrained are evident in this study. Thermal diffusivity is defined by Hillel (1982) as the change in temperature in a unit volume of soil caused by heat flowing through the volume in unit time under a unit temperature gradient. Since temperature profiles differ between drained and undrained areas, differences in thermal diffusivity must exist. In the undrained area the change in temperature is gradual, but occurs throughout the soil volume, i.e. there is a gradual rise in temperature occurring at all depths. In the drained area the temperature rise is rapid and is confined to the region adjacent to the heat source (in this case the peat surface).

The effects of evaporative cooling appeared to be restricted to the upper 10 cm. The undrained area was cooler at the surface (Figure 5), due in part to greater evaporation rates. At greater depths, however, the higher thermal conductivity of the moist, albeit cool, surface peat in the undrained area resulted in higher temperatures in undrained vs. drained, and thus more rapid thaw relative to the drained area.

F. Freezing of substrate over winter

Frost tubes were used to sample substrate frost advance over winter. On October 29, 1985, the tubes indicated presence of a small amount of surface frost at 100% of the sample points in both the 25m and 40m spacing zones in the drained area. In the undrained area 71% of the tubes indicated frost at this time. It was expected that the undrained area would take longer to freeze because of the higher heat capacity of the surface peat in this area. By the next reading on November 7, however, all tubes in the undrained area indicated a frozen zone.

Initially freezing was slightly faster in the 40 m spacing area (Figure 7). On November 7, 1985, this area had an average frost depth of 21.2 cm in hummocks and 11.3 cm in hollows. The 25 m spacing and undrained areas had frost depths of 12.4 and 15.8 cm (hummocks) and 5.7 and 9.9 cm (hollows) respectively. Slower freezing of the 25 m spacing zone and undrained area may be attributed to higher water contents and therefore higher heat capacity.

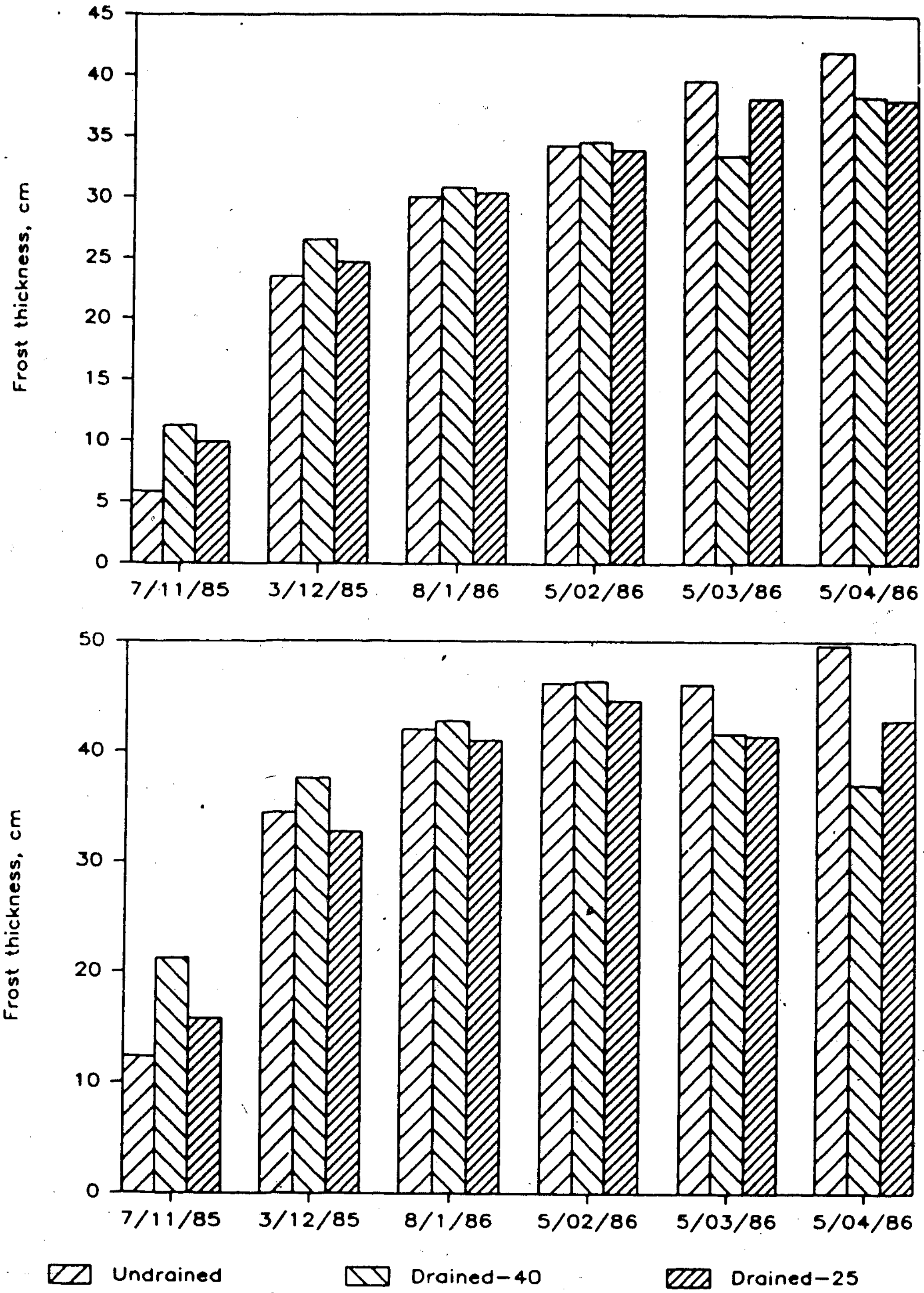

Figure 7. Thickness of the frozen layer in hollows (upper graph) and hummocks (lower graph) as indicated by frost tubes for winter 1985-1986.

Frost formation was similar in all areas for the months of December 1985 through February 1986 (Figure 7). The frozen layer in the undrained area continued to form in March and April, but in the drained areas frost thickness remained stable during this period. This suggests that differences in thermal conductivity were evident during the cooling phase as well. Higher thermal conductivity in the undrained area would allow more heat to escape from the substrate and thus cause greater frost thickness. The fact that this trend was not observed during the winter months is probably due to snow cover, which would have had an insulating effect that would be roughly equal for all areas. The greater rates of freezing in the undrained area were evident only after the snowpack had begun to melt.

Frost tubes performed adequately throughout the winter. By spring the upper frost boundary became difficult to read in several tubes due to leakage of fluorescein at the bottom of the tube. This caused the upper portion of the tube to dry out. Three more tubes were rendered inoperative by inadequately sealed outer tubes, which allowed water to pass up into the outer tube. This caused the inner tube to freeze into place, making it impossible to extract for monitoring.

G. Implications for snowmelt runoff from peatlands

The results of this study contain implications for peatland management and utilization in Northern Alberta. The timing of frost recession and snowmelt runoff are of hydrological interest. Snowmelt and associated runoff occur on average from mid to late April, when most peatlands are still frozen (Figure 8 - it should be noted that this figure is for illustration of timing of events only - the study area is only a small portion of the Saulteaux watershed). This explains the rapid snowmelt runoff from peatlands compared to more prolonged runoff from adjacent mineral soil sites where infiltration is possible (Slaughter and Kane 1979). Ryden et. al. (1980) observed lateral flow

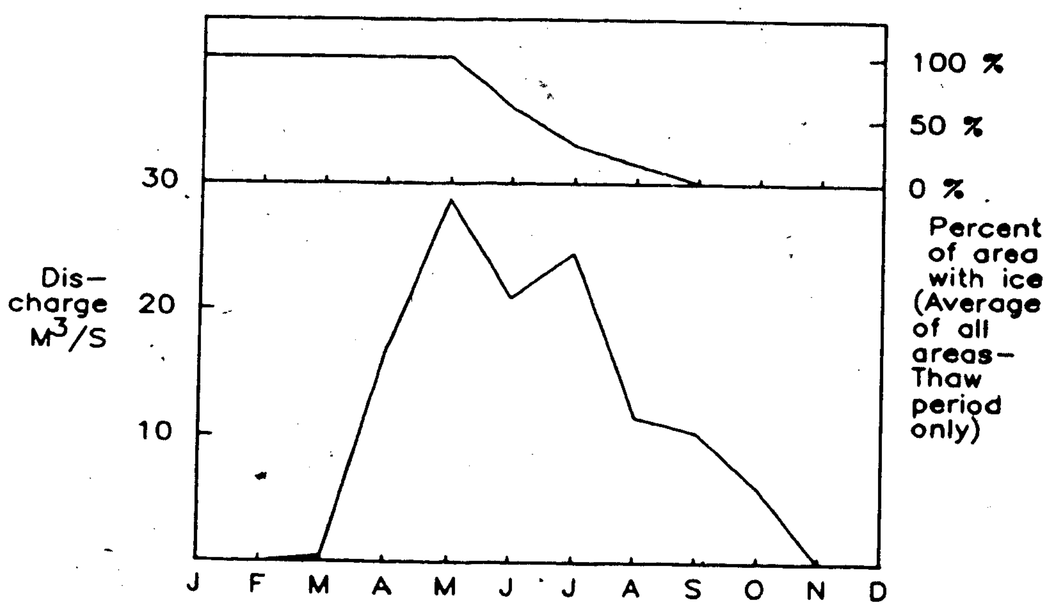

Figure 8. Study area thaw and Saulteaux River mean annual discharge.

along the frost table which then accumulated in hollows. In the case of a drained peatland this lateral flow would be stored temporarily and then continue from the hollows into ditches and subsequently out of the peatland.

H. Constraints and limitations of the study

In retrospect it is evident that several things could have been done to improve this study. A primary problem was the study was for a single site during a single year. To fully evaluate the relationships between drainage and substrate frost it would be necessary to observe the area over several years to remove the confounding effects of yearly weather differences. In addition, several areas should be examined to assure that results are not unique to a particular location.

Given the constraint of a single field season, there are several steps which could have been taken to strengthen the results of this study. It would have been desireable to find undrained control areas with saturated hydraulic conductivities similar to the pre-drainage conductivities of the 40 and 25m spacing zones. It would also have been prudent to install water wells and monitor water table levels and to measure saturated hydraulic conductivity at each sample location. Finally, a major contribution to the study would have been the measurement of peat moisture content over time and at various depths and, if possible, conduct laboratory analysis of peat thermal properties of samples from the study area. It is hoped that the constraints of this study will be taken as suggestions for further work in this area.

IV. Conclusions

The objective of this study was to evaluate the effect of drainage on freeze/thaw cycles and frost depths on a boreal wetland site and to determine if freeze/thaw cycles and frost depths were related to spacing of ditches or saturated hydraulic conductivity. The results of this study indicated that drainage altered freeze/thaw cycles and frost depths in a boreal wetland site. Because of the insulating nature of dry surface peat, drained areas remained cooler at lower depths through the summer and thus retained frozen peat layers longer than undrained areas.

The hydraulic conductivity of the peat substrate affected the thickness and structure of the frozen layer. Peats of high hydraulic conductivity had thinner frozen layers because of rapid drainage of water from the substrate. This in turn affected the timing of frost recession, because a thinner frost layer will require less heat input for thaw than a thicker frost layer.

The 25m and 40m spacing areas differed in ditch spacing as well as saturated hydraulic conductivity, therefore the separate effects of each are difficult to evaluate. The results, however, suggest that saturated hydraulic conductivity was more important than spacing distance in determining frost thickness. The thinnest frost layer in the 40 m spacing area suggested this area had less water available for freezing. On the basis of drainage intensity alone (i.e. if both areas had equal hydraulic conductivity), one would expect drainage in the 25m spacing area to be more effective than in the 40 m area because of the more intensive ditch density. Since the opposite appeared to happen, the effect of differences in ditch spacing was apparently overwhelmed by the effect of differences in hydraulic conductivity, or more precisely differences in pore size distribution implied by hydraulic conductivity differences. It should also be noted that the saturated hydraulic conductivity for the drained areas referred to in this study was measured below the water table prior to drainage, and it has been assumed that these values are representative of differences in unsaturated hydraulic conductivity above the water table after drainage. A logical next step in this study would be to verify this by measuring unsaturated hydraulic conductivity in the laboratory. One could then determine the relationship between hydraulic conductivity and water content in the unsaturated zone.

Drainage of peatlands in Northern Alberta must be approached cautiously. A consequence of drainage will likely be the alteration of summer thermal regime in the area, with surface layers warmer than the undrained condition but subject to much wider diurnal fluctuation. At lower depths temperatures will be depressed compared to the undrained condition.

Drainage under conditions of shallow humic peats of low hydraulic conductivity could lower substrate temperatures enough to induce localized permafrost. The area chosen for this study is well south of both the established limits of discontinuous permafrost and the 0 °C mean annual air temperature isotherm, yet 16.7 % of the sample points in the 25 m spacing zone retained ice through the summer and into the fall of 1985. Permafrost occurring naturally in Northern Alberta is confined mainly to peatlands, specifically where drying of surface peat occurs in summer (Brown, 1977). One would suspect that peatland drainage, which causes the drying of surface peat, could easily induce permafrost formation in this region. Further study is needed to verify specific effects of drainage on thermal regimen, for example with respect to peat type and local hydrology.

It is well known that a primary factor inhibiting tree growth in peatlands is poor root aeration. Lowering of water table levels through drainage and thus providing an aerated rooting zone is a major positive step in improving growth. This study has concentrated on one possible consequence of drainage, changes to thermal regime. It is not known, however, if these changes will have a significant effect on tree growth. Some possible negative effects might be: (1) creation of a frozen layer near the surface which would inhibit root penetration; (2) reduced temperatures which would slow metabolic activity; (3) wide diurnal temperature fluctuation at the surface which could be detrimental for seedling survival because of increased risk of both scalding and frost damage. The significance of alterations to thermal regime caused by drainage needs to be investigated relative to silvicultural and horticultural requirements of species planned for management on peatlands.

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