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LE GEL DE TERRAIN TOURBEUX

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SOMMAIRE

On évalue à plus d'un million de kilomètres carrés l'étendue de terrain tourbeux qui couvrent le Canada, de son point le plus au sud jusqu'à l'Arctique, dans la zone à pergélisol continu. Le développement rapide du Nord a donné lieu à de nombreuses études de ces terrains par des hommes de sciences et des ingénieurs.

Le gel de terrain tourbeux affecte de nombreuses activités géo-
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NATIONAL RESEARCH COUNCIL OF CANADA

DIVISION OF BUILDING RESEARCH

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SUMMARY

It is estimated that peatland covers more than one million square kilometres of Canada, extending from the southernmost part of the country to the Arctic in the continuous permafrost zone. This terrain has been subjected to increasing scientific and engineering study as northern development progresses.

Many geotechnical engineering activities in Canada are affected by the freezing of peatland; the operation of off-road vehicles on peat terrain, the construction of oil and gas pipelines, and the erection of temporary structures all require information on the freezing process in peat terrain. The present paper is intended to give engineers and other workers a practical appreciation of the rate of freezing and thawing, depth of frost penetration and thaw, and the influence of climate and terrain on these processes. It is based on information available in the literature and on field observations at two sites, one the Mer Bleue peat bog near Ottawa in the zone of seasonal freezing, the other at Thompson, Manitoba, in the middle of the discontinuous permafrost zone.

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Vast areas of Canada are covered by extensive peat deposits. No detailed mapping of the distribution of this type of terrain has been undertaken, but its area is estimated to be at least 1,300,000 sq km (500,000 square miles) extending from the southernmost parts of Canada to the Arctic in the continuous permafrost zone. In recent years this unique terrain, variously referred to as bogland, muskeg, organic terrain, and peatland, has been subjected to increasing scientific and engineering study as northern development progresses. In 1969 the Muskeg Engineering Handbook was published to bring together this knowledge and experience (MacFarlane, 1969).

Many geotechnical engineering activities in Canada are affected by the freezing of peatland; the operation of off-road vehicles on peat terrain, the construction of oil and gas pipelines, and the erection of temporary structures all require information on the freezing process in peat terrain. These problems associated with freezing have received minimum treatment in the literature because little work has been carried out in this particular field in Canada, although there has been considerable interest in the subject in the U.S.S.R. (Deryugin and others 1969; Romanova and Ivanov 1969). This paper is intended to give engineers and other workers a practical appreciation of the rate of freezing and thawing, depth of frost penetration and thaw, and the influence of climate and terrain on these processes. It is based on information in the literature and on field observations from two sites, one a bog (Mer Bleue) near Ottawa in the zone of seasonal freezing, the other at Thompson, Manitoba, located in the middle of the discontinuous permafrost zone (Figure 1).

Before proceeding it may be useful to consider terminology, particularly with regard to the designation of the terrain under consideration. The term "peatland" is preferred to the others mentioned above. It is the peat component that makes this type of terrain unique and thus different from all other types. The term also enjoys widespread international use in earth sciences and geotechnical literature.

The other three do not. Although "muskeg" is used widely in North American geotechnical literature, it is an awkward term which has never been defined satisfactorily and has not been accepted by biologists and earth scientists on this continent or abroad. "Bogland" is used in the Russian literature, but is too restrictive because bogs comprise only one type of peatland. "Organic terrain" is unsuitable because all vegetated terrain on the earth's surface includes organic matter.

Three categories of peatland are described in the Muskeg Engineering Handbook relative to the distribution of permafrost. The first, "seasonally frozen", is found south of the permafrost boundary where the Mer Bleue peat bog is located. Peat ranges in thickness from 30 cm to more than 6 m. As the name implies, the surface-frozen layer is entirely seasonal. The depth of frost penetration or thickness of the frozen layer is extremely variable, depending upon climatic factors such as air temperature, solar radiation, and snow cover.

The second category is peatland in the discontinuous permafrost zone where Thompson is located. Observed peat thickness varies from about 30 cm to 3 to 5 m in rock basins and palsas. In the southern fringe of this zone permafrost occurs in scattered islands a few square metres to several hectares in size and is confined mainly to peatland. Northward it becomes increasingly widespread and occurs in other terrain types also. Where permafrost exists in peatlands the depth to the permafrost table is usually less than 1.5 m. The active layer usually extends to the permafrost table, but a variation of this category may occur where an unfrozen layer exists through the winter between the seasonally frozen peat and the permafrost.

Peatlands in the continuous permafrost zone comprise the third category. Observations have shown that these peat deposits are generally relatively shallow, their thickness seldom exceeding 1 m or a few feet. Permafrost occurs everywhere beneath the ground surface, including all peatland. The active layer generally varies in thickness from about 0.3 to 1 m in the continuous zone depending on local climatic and terrain conditions and it almost invariably extends to the permafrost table. The active layer is thinnest in peatlands, an annual depth of thaw of only 15 cm being observed in the Arctic Archipelago.

PEATLANDS - CLIMATE AND TERRAIN

Peat deposits differ from mineral soil in that they are composed entirely of vegetal material. Their formation and characteristics and their freezing regime are greatly influenced by climatic and terrain factors (Brown, 1966; Brown, 1970; Tyrtikov, 1959). The most extensive peat deposits in Canada occur in the discontinuous permafrost zone. Vegetal source material in the boreal forest, both woody and non-woody, is available in virtually the same quantities as south of the permafrost region, but decomposition rates are generally slower because of the colder climate. This results in peat accumulations of several decimetres per thousand years.* The top layer, several centimetres to a maximum of about 30 m, consists of combinations of living mosses, lichens and sedges. This living material grades downward into dead, partially decomposed peat of varying thickness and composition overlying the mineral soil. In contrast with peatland the ground surface of mineral soil terrain is often comprised of a layer of organic material consisting mainly of leaf and needle mould and some moss and lichen overlying the mineral horizons.

Peatlands are usually flat, but can develop on gentle slopes, and this has been observed, for example, in Newfoundland where cool marine climatic conditions prevail. The surface of peat deposits confined in rock basins often displays a convex form, being slightly domed in the centre.

Considerable microrelief occurs in peatlands in the form of peat mounds, ridges and plateaux rising generally 1 to 1.5 m above the general surface. These are particularly prevalent in the permafrost region owing to ground ice build-up.

The living vegetation associated with peatlands differs significantly from that of mineral soil terrain. Tree growth is almost entirely coniferous, black spruce being the dominant species in Canada. Individual trees are generally more stunted and scattered than the same species on mineral soil terrain because of poor drainage and high acidity. Undergrowth is frequently deciduous,

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* Rates ranging from 30 to 70 cm per thousand years have been determined by radio carbon dating methods carried out by the Geological Survey of Canada.

consisting of willow, alder shrubs and ground birch. Lower plants rooting in the moss cover grow in profusion, dominated by the bushy Labrador tea. The ground cover itself consists of combinations of mosses, lichens and sedges. Sphagnums and other mosses predominate and usually form a continuous and frequently hummocky carpet with patches of lichen growing over them. Sedges are more common in wetter parts of peatlands.

Attempts have been made to classify peatland and peat (Lacate 1969, MacFarlane 1958, Radforth 1952). Some success has been achieved, but difficulties arise in relating living vegetation patterns to underlying peat characteristics. Despite some claims, correlations do not exist because the peat was formed from vegetation existing centuries or millenia ago. Changes in drainage and other surface conditions can cause corresponding changes in the living vegetation, resulting in its bearing little relation to the dead matter below. The classification of peat itself is very difficult because of the innumerable combinations of woody and non-woody matter having fibrous and granular texture. This categorization is further complicated for engineering purposes by freezing where the quantity and distribution of ice in the peat becomes a factor. This can vary greatly, depending on temperature and particularly on the unusual water retention properties of the peat which because of its structure can hold up to fifteen times its weight of water.

The various types of plant cover on peat influence the surface heat exchange processes that determine the rate of freezing and thawing and ground temperature regime. Vegetation influences the depth and density of the snow that accumulates and thus, indirectly, changes the amount of heat flowing into or out of the ground. The uneven surface of peatland has an effect on convection and evaporation heat exchange because surface irregularities change surface heat and moisture transfer coefficients. The extreme variability of vegetative covers and the complexity of the surface heat exchange processes make it difficult to obtain reliable values for the components of the surface heat balance for peatland.

A characteristic upon which peat formation depends is poor drainage conditions. Surface water from rain or snowmelt remains on or close to the surface, producing flooded conditions. Consequently, much of the heat from solar radiation is expended in evaporation and is not available for warming the soil. Peat soils, with generally high moisture contents, have high latent heats of fusion, a factor which is important in freezing and thawing processes.

Soil-water conditions are governed by precipitation, evapotranspiration, microrelief and the physical characteristics of the Sphagnum or other cover. The moisture content of the porous cover of Sphagnum on peat bogs responds rapidly to precipitation. Water can move by convection in water-filled pores, by gravity along the interface between the surface cover and underlying saturated peat, and by vapour movement under unsaturated conditions. In contrast, water in underlying decomposed peat is frequently almost immobile, with nearly constant moisture content in the soil from year to year.

The highly variable water content and porous structure of Sphagnum and other surface covers makes the process of heat transfer in the surface layers much more complicated than in the underlying decomposed peat. In saturated peat, heat transfer takes place almost entirely by conduction. In porous Sphagnum, however, heat can be transferred by conduction, convection or transfer of water vapour. Romanov (1961) estimates that in Sphagnum, at low moisture contents,up to 20 per cent of the total heat is transferred by the diffusion of water vapour; at high moisture contents heat transfer is entirely by conduction and convection (heat transfer in water-filled pores).

The thermal properties of peat depend on moisture content to a far greater extent than for most mineral soils. Variations in porosity, plant structure and temperature also influence thermal properties but, judging from the limited information in the literature, these effects are secondary. Figure 2 shows the variation of thermal conductivity with moisture content for peat, silt, clay and sand. The range in thermal conductivities for the sand and silt clay is for low and high values of dry density and for low and high values of moisture content (Brown 1964). The range in values for peat was obtained from the literature (Williams 1966, Romanov 1961, Lachenbruch 1959, Kersten 1949). The thermal conductivity of unfrozen peat at high moisture contents is in the range of the lowest reported values for mineral soils. Thermal conductivities reported for Sphagnum at the most common range of moisture content of 30 to 60 per cent (Romanov 1961) are much lower than for mineral soils. The thermal conductivity of dry Sphagnum is probably an order of magnitude lower than the lowest values for mineral soils, comparable to that of medium density snow (Williams 1968).

The volumetric heat capacity (Cv) of peat soils depends almost entirely on moisture content. Values range from about 0.15 cal/cm^3/°C for unfrozen Sphagnum of low water content to about 1.0 cal/cm^3/°C for unfrozen saturated peat. Because peat soils have relatively low thermal conductivities and high volumetric heat capacities their thermal diffusivities k/Cv, ranging from 0.0005 to 0.0015 cm^2/sec, are low compared with those for mineral soils (0.002 to 0.016 cm^2/sec) (Geiger 1965).

The depth to which diurnal and annual temperature waves penetrate peat is much less than that for inorganic soils. Figure 3 illustrates the annual variation of ground temperatures at various depths in peat soil (Williams 1968) and nearby silt-clay soil (Gold 1963). The depth to which ground temperatures are affected by annual changes in air temperature is about 4 m for peat compared with about 7 m for mineral soil. In this particular case the mean annual soil temperature and the amplitude of the annual surface temperature wave were considerably lower for peat soil.

FREEZING OF PEATLAND IN THE SEASONALLY FROZEN ZONE

Frost depth was measured during two winter seasons at the Mer Bleue peat bog, Ottawa, to provide quantitative information on ground heat loss and variations in the depth of frost penetration due to terrain conditions. The principal surface vegetation is Sphagnum about 15 to 30 cm thick, covering peat deposits varying from 1.5 to 5 m in depth. Other forms of vegetation frequently associated with peat bog such as blueberry bushes, spruce and tamarack are present (Figure 4). The growing Sphagnum and other vegetation are typical of bog vegetation occurring at higher latitudes.

Observations of Freezing Regime

Frost depth was measured during the 1968-69 and 1969-70 seasons using frost depth gauges (Gandahl 1963) consisting of 1.2-m, 1.3-cm O.D. plastic tubes filled with methylene blue dye and placed inside 1.2-m, 2.2-cm O.D. plastic tubes. The tubes were installed vertically in the ground with about 30 cm protruding above the soil surface. The depth to the 0°C isotherm was measured by removing the inner tube and noting the depth at which the dye changed colour. (Dye changes colour when the solution freezes.) The 0°C isotherm is a good indicator of frost depth; most of the water in peat soils freezes at 0°C because the pores are large and the mineral content small (Romanov 1961). Figure 5 shows the snow depth and thickness of frozen layers during two winter seasons for a typical hummock and depression.

Ten frost depth gauges were installed in 3-m circular patterns around typical Sphagnum hummock-depression relief. Five additional gauges were installed at other locations about 45 m away from the site (where detailed measurements were made). Frost depths were checked by hand augering. Check levels taken when the frost gauges were first installed and later in the winter season showed that there was no vertical movement of the gauges from frost heave. Figure 6 shows the thickness of snow cover and the thickness of the frozen layer at different times during the two winter seasons at ten observation points within a 3-m circle.

These observations illustrate the importance of snow cover in reducing the rate and maximum depth of freezing. The hollows between hummocks can be filled by drifting snow even when there is light snowfall. Once the hollows are filled, the thermal resistance of the snow in the hollows and Sphagnum in the hummocks creates an effective insulating layer. The time when the snow cover forms is also important. The deep snow cover early in the winter of 1968-69 prevented significant frost penetration in the depressions for the rest of the winter.

Vegetation growing on hummocks tends to increase the amount of snow trapped and hence decrease the amount of frost penetration. In a random survey of maximum frost depth in the 1967-68 season the maximum depth of freezing for hummocks with no vegetation was 40 to 42 cm compared to 33 to 37 cm for hummocks with medium bushes and 31 to 33 cm for hummocks with heavy bushes.

The rapid freezing of hummocks early in the winter season is a marked characteristic of the freezing pattern of peatland. The hummocks freeze quite rapidly because of their low water content and relatively low latent heat of fusion. The large exposed surface area of the hummocks also increases heat loss by long wave radiation and convection leading to rapid freezing (Romanov 1961).

The drainage of water from hummocks into depressions affects the freezing pattern. According to field observations in the U.S.S.R. (Chechkin 1965), hummocks with steep slopes ensure good drainage of water and freeze more rapidly than hummocks with shallow slopes. The height of the water table and the amount of rain in the fall also have a marked influence on their rate of freezing.

The movement of water from the centre of the bog delays freezing in depressions near its borders. Open water observed near the edge of the Mer Bleue bog during most of the 1968-69 winter is, apparently, not an uncommon feature of peat bogs (Romanov 1961).

The ice growth at all stations was less than 1 cm during the period from 5 to 26 February 1969. The heat flow through the snow cover and frozen layers was calculated for this period, assuming steady-state conditions for the fifteen stations where frost depth was measured. The mean temperature of the snow surface was assumed equal to air temperature measured in a Stevenson screen located at the site. The following average values of thermal conductivity were used in the calculations:

snow - 0.0003 cal/cm-sec/°C,
frozen Sphagnum - 0.0004 cal/cm-sec/°C,
frozen saturated peat - 0.004 cal/cm-sec/°C.

Table I lists the calculated values for heat flow at the fifteen stations.

These calculated values for heat flow through the snow and frozen layers were compared with calculated values for heat flow through the underlying, unfrozen peat. Steady-state conditions, a thermal conductivity of unfrozen peat of 0.001 cal/cm-sec/°C, and an average temperature gradient of 0.06°C/cm were assumed. The temperature gradient was obtained from ground temperatures measured at 10-, 30- and 100-cm depth using mercury-in-steel thermometers connected to a 3-pen recording thermograph. The calculated total heat flow of 108 cal/sq cm for the three-week period agrees with the average value for heat flow through the snow and frozen layers. These results show that heat flow from underlying peat is sufficient to limit frost depth penetration in peat soils once snow has filled the hollows between Sphagnum hummocks. (During February 1970 the thickness of the frozen layer decreased, although air temperatures were still well below freezing.)

Frozen peat under hummocks thaws much slower than the frozen layer in hollows (Figure 7 illustrates the slow thawing of frozen peat under a hummock). The low thermal conductivity of the Sphagnum in the hummock retards the melting of the frozen layer. Snow meltwater, collecting in the depressions, is one factor causing fairly rapid thawing of the frozen peat.

The duration of the thaw period varies greatly from year to year. In 1969 all the ice had melted by 15 April, coinciding with the final melting of the snow cover. In 1968 the frozen layer under the hummocks persisted until about the second week in June. In the previous year ice persisted until late June. The presence of ice under peat hummocks throughout the summer is a common feature in permafrost areas. In eastern Siberia the permafrost table is often a mirror image of the hummock surface relief (Kudryavtsev 1959).

Maximum Depth of Frost Penetration

The maximum depth of seasonal freezing is often required for engineering purposes. As there is little information in the literature on this aspect of peat soils values of maximum depth of freezing for different freezing indices were calculated using a procedure based on the modified Berggren equation (U.S. Corps of Engineers, 1966). Average values for the thermal conductivity of peat and Sphagnum were assumed.

Figure 8 shows the maximum calculated depth of freezing of saturated peat with and without a 30-cm layer of Sphagnum. The modified design curve and frost penetration for silt-clay and sandy soils without a snow cover (Brown 1964) are also shown. Field measurements available for the maximum depth frost penetration in peatland for different values of freezing degree days are plotted on the graph.

The mean annual air temperatures corresponding to various accumulated freezing days are also shown on Figure 8. The freezing index (FI) was related to mean annual temperature by assuming that mean monthly air temperatures follow a sine curve. The following equation shows this relation for mean annual temperatures less than 0°C (32°F):

$$FI = (32-\overline{T}) \left(\frac{P}{2} + \frac{P}{\pi} \sin^{-1} \left(\frac{32-\overline{T}}{A}\right)\right) + \frac{AP}{\pi} \sqrt{1 - \left(\frac{32-\overline{T}}{A}\right)^2}$$

where

$\overline{T}$ = Mean annual temperature (°F)

P = period 365 days

A = amplitude of Sine Curve

$$= \frac{\text{Mean temperature July + Mean temperature January (°F)}}{2}$$

An equation of the same form can be developed for mean annual air temperatures greater than 0°C. This relation can be used to obtain the freezing index directly from mean annual air temperature data readily available in meteorological publications. It is specially useful for calculating the depth of freezing in peatlands because it changes slowly in comparison with changes in FI (Figure 8).

These calculations show that the maximum depth of frost penetration in peat is much lower than for all other soil types. The maximum depth for saturated peat covered by a 30-cm layer of Sphagnum is probably under 60 cm for most regions in Canada south of the discontinuous permafrost zone. If the Sphagnum layer is removed the maximum depth for saturated peat without snow cover is probably 100 cm or less. The average maximum depth of freezing observed for peatland in different regions of the U.S.S.R. (Checkhin 1965) is less than the values calculated for a Sphagnum layer on saturated peat (Figure 8). The actual depth of freezing will generally be less than calculated values because snow usually fills the hollows and covers the hummocks. Variations of freezing depth occur from one peat area to another and even within individual bogs and swamps. In general, however, the depth of freezing of depressions or hollows appears quite predictable and the variations resulting from differences in topography, vegetation cover and water content appear to fall within well defined limits (Chechkin 1969).

FREEZING OF PEATLAND IN THE DISCONTINUOUS PERMAFROST ZONE

Investigators are primarily interested in the maximum depth of thaw that will take place in permafrost areas under given climatic and terrain conditions. This is in contrast with the seasonally frozen zone where the maximum depth of freezing is the prime consideration (Deryugin and others 1969). The same factors of snow cover, soil thermal properties and surface heat exchange, however, also affect the thawing process. Observations on seasonal freezing of peatland in a permafrost zone are presented in this paper to show the variations that can occur at a site because of variation in climatic and terrain factors.

Description of Sites and Measurements

Observations on seasonal freezing of peatland in a permafrost zone have been undertaken at Thompson, Manitoba, since 1968. This community is about 640 km north of Winnipeg in north central Manitoba in the middle of the discontinuous zone. Permafrost occurs in scattered islands varying in extent from a few square metres to several hectares and in thickness from about 1 to 15 m or greater, averaging between 2.4 and 4.5 m (Johnston et al 1963). The permafrost table is generally encountered anywhere from about 0.5 to 2 m below the ground surface. Much ice, primarily in the form of horizontal lenses up to 20 cm thick (the average thickness being less than 2.5 cm), is found throughout the frozen glaciolacustrine silts and clays underlying the area. Temperatures in the permafrost vary from about -0.5 to 0°C.

These observations are one phase of a program of investigation of microclimatic and terrain factors affecting the distribution of permafrost in the discontinuous zone. Four sites located within a distance of a few hundred metres of each other, two with permafrost and two with none, were selected for study. Regular observations are being carried out of air temperature, precipitation, snow depth and density, wind speed at the 1.8-m level, net radiation, ground heat flow at the 30-cm depth, ground temperatures at 12 depths to 7.5 m, and annual depth of ground freezing and thawing. The last two measurements are made with methylene blue frost gauges installed to a depth of 1.5 m in the two permafrost sites and 3 m in the two sites with no permafrost.

Soil profiles of the four sites are shown in Figure 9. Site A supports dense spruce (Figure 10a) growing on a 10-cm thick layer of feather moss and forest litter overlying brown clayey soil (grain size analysis at a depth of 15 to 23 cm: clay - 59 per cent, silt - 39 per cent, sand - 2 per cent). There is no peat at this site and no permafrost.

Site B supports open stunted spruce (Figure 10b) growing on a 10-cm thick layer of Sphagnum over 10 cm of peat, 23 cm of black organic clay (31 per cent organic content by weight) overlying brown clayey soil (grain size analysis at a depth of 30 to 53 cm: clay - 77 per cent, silt - 22 per cent, sand - 1 per cent). There is no permafrost at this site.

Site C supports open stunted spruce similar to that at Site B (Figure 11a) growing on an 18-cm thick layer of Sphagnum over 18 cm of peat, 10 cm of black organic clay (26 per cent organic content by weight) overlying brown clayey soil (grain size analysis at a depth of 35 to 60 cm: clay - 85 per cent, silt - 14 per cent, sand - 1 per cent). There is permafrost at this site. The permafrost table in 1968 was at a depth of about 48 cm almost coincident with the base of the organic soil, and the permafrost is about 2.7 m thick.

Site D supports dense spruce, similar to that at Site A (Figure 11b) growing on a 13-cm thick layer of Sphagnum over 18 cm of black organic silt (77 per cent organic content by weight) overlying brown clayey soil. Grain size analysis of the organic silt is: clay - 28 per cent, silt - 62 per cent, sand - 10 per cent; in the brown clay at 33 to 48 cm it is: clay - 87 per cent, silt - 12 per cent, sand - 1 per cent. There is no peat at this site similar to that at Sites B and C, but the organic silt has a very high organic content. The permafrost table is at a depth of about 75 cm and the permafrost exceeds 6 m in thickness.

Ground Freezing Regime

Frost depth-time graphs are shown for each site for three complete freezing seasons (1968-69, 1969-70, 1970-71) in Figure 12. At the two sites with no permafrost seasonal frost penetrates to its maximum depth in late winter, followed by thawing from the surface downward in spring. At the sites with permafrost seasonal thawing progresses downward each summer, reaching the permafrost table in the early autumn. This is followed by seasonal freezing from the surface downward to the permafrost table.

Several patterns are evident on the basis of observations over three freezing seasons, for there are significant differences in the freezing and thawing regimes at the four sites. Site A experiences considerably deeper frost penetration and a longer period of seasonal freezing each year than Site B. Both sites are without permafrost and only Site B can be considered a peatland site. Both Sites C and D are in permafrost areas and seasonal frost penetrates to the permafrost table each year. The active layer is significantly thicker at Site C and changes from year to year. It appears to remain constant at Site D.

The rate of frost penetration is more uniform at Sites A and B where there is no permafrost than at Sites C and D where permafrost complicates the seasonal freezing of the ground. Frost penetration for all four sites in Table II is presented for the three freezing seasons. Depths and penetration rates are greater at Site A than at Site B; penetration rates at both sites are generally two to three times greater in the first part of the winter than in the latter part, although brief periods at lower rates occur at the very beginning. Thawing of the ground begins in the early spring at about the time the frost penetrates almost to its maximum depth. Further ground freezing of a few centimetres at most does occur during the following two-to eight-week period but its effect in the total freezing regime is minimal. The last remnant of frozen ground disappears in late summer about one month later at Site A than at Site B. At both sites it was located at the bottom of the frozen layer, indicating little or no thawing from beneath.

The situation is somewhat different at Sites C and D where permafrost exists, although that at Site C is more marginal than that at Site D. At Site C the temperature of the permafrost is about -0.1°C and thus is very sensitive to any disturbance. Three years of weekly observations have resulted in a drop in the permafrost table from the original level of 48 cm to 1.35 to 1.5 m. At Site D, where the temperature of the permafrost is slightly colder at about -0.5°C, the permafrost table has remained steady at a depth of about 0.75 m. Although the frost gauges at both sites were installed in the fall of 1968 the methylene blue liquid in the bottom few decimetres of the gauges in the permafrost did not freeze until mid-winter.

Frost penetration rates are generally more uneven at Sites C and D than at the sites with no permafrost. This was particularly true in 1968-69 and 1969-70 when penetration rates were rapid during the first two or three weeks, levelled off in the late fall, and then increased as the seasonal freezing plane approached the permafrost table. At Site D the seasonal frost reached the permafrost table at the end of December 1969-70, the most severe of the three freezing seasons. This season produced the deepest frost penetration at the non-permafrost sites. At Site C seasonal frost reached the permafrost table in mid-February 1969-70. Ground thawing of the active layer begins at about the same time as for the two sites with no permafrost.

Observations on Climatic and Terrain Factors

Observations of climatic and terrain factors are tabulated in Table III. Net radiation observations are not tabulated because they were only commenced on a continuing basis during the summer of 1971. Wind velocities were measured at the four sites through the freezing seasons of 1969-70 and 1970-71. Average wind speeds are very low, less than 1.6 km/hr compared with an average of approximately 16 km/hr at the standard meteorological installation 12 m above ground level at the Thompson airport.

The mean air temperature for the eight-month winter period of each of the three freezing seasons was slightly higher at Site A than at Site B, both sites with no permafrost. The same was true for the four summer months. At the permafrost sites the mean winter air temperature was generally lower at Site C than at Site D. The mean summer temperature was consistently lower at Site D, the lowest of the four. The three-year average encompassing the three freezing seasons, October 1968 to September 1971, was highest at Site A and -0.3 C deg less at the other three sites, which were virtually the same. These values are similar to the value of -3.6°C for the same period at the meteorological station at Thompson.

Mean ground surface temperatures followed somewhat the same pattern during the winter. Site A had generally the highest values of the four sites and Site D the lowest. During the summer Site A was consistently low and Site B, the other site with no permafrost, high. The three-year average encompassing the three freezing seasons, October 1968 to September 1971, was highest at Site B and lowest at Site D.

Ground heat flow measurements are available through the two freezing seasons 1969-70 and 1970-71. There is considerable difference in average values for heat flow for the two years of observations. During both winter seasons Site D had the lowest value (less heat lost to the atmosphere). During both summers Site C had the highest positive values (heat flow from the atmosphere to the ground).

The surface organic layer is an important factor in the freezing regime of the ground. The organic profile at each site has been described and is tabulated in Table III. The organic layer is thicker at Sites B and C, 43 and 45 cm respectively and thinner at Sites A and D, 10 and 30 cm respectively.

Snow cover is one of the most important factors influencing the freezing of the ground. The mean thickness of the snow layer has been consistently highest at Sites B and C, where tree growth is sparse. Site D has the lowest snow accumulation and the densest, tallest tree growth. The pattern of snow density was less consistent, although values tended to be slightly lower at Site D than at the other sites.

Discussion of Results

The three major factors determining the freezing and thawing pattern at a site are:

(1) heat exchange at the surface by radiation, convection and evaporation,

(2) depth and density of snow cover,

(3) thermal properties of the soil.

The assessment of the relative influence of these factors is complicated by their close interrelationship. Often virtually unmeasurable changes in one of the factors or in the weather and soil properties upon which they depend can cause significant differences in soil freezing or thawing. A discussion of the observations can, therefore, only illustrate in a qualitative way the importance of various variables, particularly that of the surface organic layers, in the freezing and thawing of ground at these four sites.

There are not sufficient data on radiation, wind velocity and surface moisture to allow discussion of the observed freezing and thawing patterns from a surface energy exchange point of view. In winter, differences in evaporative and convective heat losses at the four sites tend to be undetectable; wind velocities are extremely low and all sites have generally the same surface conditions.

Differences in net radiation may not be very great because the greater amount of solar radiation received at exposed sites could well be offset by higher long-wave radiation losses to the atmosphere. The fact that average air temperatures during the winter are almost identical at all sites tends to support this conclusion. In summer, differences in surface heat balance depend to a great extent on surface moisture conditions which control evaporation. The relative importance of evaporation and other components of surface energy exchange during the thaw period will become clearer with continuing measurements over periods of several years.

During the winter, snow cover largely controls the amount of heat lost from the ground and subsequent ground freezing at the four sites. Figure 13 shows average calculated heat flow for each site for each snow season (1 November to 30 April). These calculations were made by assuming steady-state conditions, and an average snow surface temperature equal to the average air temperature. The average temperature gradient through the snow cover was assumed to be equal to the difference between average air and ground temperatures divided by the average depth of the snow. The thermal conductivity of the snow was calculated from snow density values (Williams and Gold 1958).

The results of the calculations shown in Figure 13 indicate that there was considerable variation in heat loss through the snow cover from year to year. Heat loss was highest at all sites during 1968-69 and lowest during 1970-71. On the average, Site A showed the largest heat loss; Sites B and C showed almost identical heat losses; and Site D showed the lowest average heat loss.

The calculated values for heat loss through the snow cover were checked by comparing them with heat loss measured at the 30-cm depth. The total heat loss for two winter seasons (1969-70, 1970-71), measured with the heat meters, is about 15 per cent higher than the estimated heat loss through the snow cover at Sites A, B and C (Figure 13). Measured heat loss at Site D is slightly lower than calculated heat loss through the snow cover. Considering the uncertainties associated with measurement of heat flow by means of soil heat meters and the fact that the heat meters were not placed at the surface, the agreement is quite good.

Differences in heat loss through the snow cover for the four sites are explained by differences in snow depth and density. Sites B and C have comparable average snow depths and densities. Site A has a greater snow depth than Site D, but higher snow density offsets to some extent the greater snow depth. If ground temperature had been the same at both sites, heat loss from Site D would have been only slightly higher than heat loss from Site A. The difference in heat flow between Sites A and D is thus primarily caused by differences in ground surface temperature.

The thermal properties of the surface soil layer have a significant influence on heat transfer to the surface. Frozen peat and Sphagnum in the upper 30-cm layer at Sites B and C should have thermal conductivities ranging from 0.003 to 0.005 cal/cm/sec°C at high moisture contents. These values compare with the range for frozen silt clay (Figure 2) at average moisture content, the most likely conditions at Site A. Consequently, average ground temperatures at the three sites are quite similar. Deviations from assumed thermal conductivities and differences in thermal gradient and snow depth and density probably account for differences in average heat loss during the winter. Site D is a drier site than Sites B and C and if the organic layer at Site D has a markedly lower moisture content the thermal conductivity of the frozen surface organic layer would presumably be lower. This combined with the lower subsurface ground temperatures associated with this permafrost site may account for the lower ground temperatures and total heat loss.

The heat loss from the ground surface in winter results in the cooling and freezing of the soil. As large amounts of heat must be extracted to freeze water in soil, the depth and rate of frost penetration depend largely on the amount of moisture present. Site A, the driest site, has a greater depth of frost penetration than Site B, which is poorly drained. The seasonal fluctuations in the rate of freezing of the first few centimetres of soil at Sites C and D (Table II) are probably caused by variations in the moisture content of the surface organic layers just prior to ground freezing.

The amount of thawing during the summer determines not only the rate at which seasonal frozen ground melts but also possible degradation of permafrost. If the heat gained by the ground in summer exceeds the heat lost in winter, the surplus heat will warm the ground at non-permafrost sites and melt permafrost at permafrost sites. The only quantitative information on the heat gained by the ground during the summer is the measurement of heat flow at the 30-cm depth at each site.

These values for summer heat gain must be treated with caution. There is no way of checking their reliability as there is for the winter period when ground heat losses can be compared with heat flow through the snow cover. The heat flow meters may be less satisfactory in unfrozen soil because they cannot measure any transfer of heat by vapour or ground water movement. Even greater errors can also be expected because the thermal conductivity of the unfrozen soil may be much less than that of the heat meters themselves. Philips (1961) shows theoretically that errors of up to +24 per cent can occur if the thermal conductivity of the surrounding medium is considerably lower than that of the heat meters. Although the absolute values of heat measured by the meters may be uncertain, therefore, they should be representative of the relative heat gained by the ground at the four sites.

Figure 13 shows that the measured heat flow into the ground (total for two summers, 1970 and 1971) was much greater than the heat lost from the ground during corresponding winter periods (1969-70 and 1970-71). The heat gained at Sites A, B and D was about the same; that gained at Site C exceeded it by about 1,000 calories. This much surplus heat would melt about 35 cm of permafrost (at a volumetric moisture content of 35 per cent), roughly the amount of permafrost thawing that took place over the two-year period.

What would cause Site C to have such a large surplus of heat compared with the other sites? Although the answer to this question must be speculative because of lack of data, an explanation can be attempted by comparing the surface heat balance at Site C with that at Site B, which has similar exposure and soil conditions. Let us assume in making this comparison that surface disturbance resulted in ponding of water at Site C so that the surface organic layers were completely saturated, whereas at Site B the surface layers of Sphagnum remained unsaturated. The surface heat balance during the summer period can be represented by the following simple equation

$$Q_s = Q_{sw} - Q_e + Q_c - Q_{lw}$$

where

Q_s = heat flow into ground

Q_{sw} = heat gained from solar radiation

Q_e = heat lost by evaporation

Q_c = heat gained by convection

Q_{lw} = heat lost by long-wave radiation.

The heat gained by short-wave radiation would probably be about the same at both sites since they have similar exposure and surface conditions. The heat used in evaporation should also be nearly the same. Although the Sphagnum at Site B is assumed to be unsaturated, evaporation will not be limited by the supply of water because Sphagnum acts as a wick, providing ample water to the surface. Both Q_c and Q_{lw} depend on air surface temperature differences; as air temperatures are about equal at the two sites the surface temperature should be significantly different if there is to be a difference in the amount of heat gained by the ground (Q_s) over the three summer periods 1969, 1970, 1971. Figure 13 shows that the average surface ground temperature for these three periods was significantly lower at Site C than at Site B. This much difference in surface temperature is sufficient to change Q_{lw} and Q_c appreciably (Geiger 1965). As the heat gained by convection would be greater and the heat lost by long-wave radiation less at Site C than at Site B, because of the lower surface temperature, the net heat available for warming the ground would be appreciably greater over the three-year period. This additional heat can be transferred more readily into the ground at Site C, even with lower surface temperatures, because the thermal conductivity of saturated Sphagnum is considerably higher than that of unsaturated Sphagnum. The foregoing analysis suggests that ponding of water at Site C affected the surface heat balance sufficiently to cause significant changes in the total heat gained at the site over the three-year summer period. Other factors such as the lateral inflow of heat by movement of surface water and manmade disturbance affecting surface heat balance may be important.

CONCLUSIONS

Several general conclusions can be drawn from review of the literature on the freezing of peatland and the field studies at Ottawa and Thompson described in this paper.

1. The term "peatland" is preferred terminology over the other terms used to designate this type of terrain.

2. There is an inherent difficulty in classifying peat because the living vegetal surface cannot be related consistently to the underlying peat formed centuries earlier, frequently in different environmental conditions. This is an important factor in assessing freezing problems because of the necessity of knowing the characteristics of the peat.

3. The thermal regime of peatlands is different from that of mineral soil terrain owing to vegetation, microrelief, high water content, and thermal properties. It is particularly difficult to deal with heat conduction in porous Sphagnum because an appreciable percentage of the heat transfer is by non-conductive processes. The thermal conductivity of peat is generally much lower that that of mineral soil. Dry Sphagnum has thermal conductivity values about an order of magnitude below the lowest value for mineral soil. These unusual features of peatland lead to lower mean annual ground temperatures, a significant factor in the distribution of permafrost.

4. The low thermal conductivity values of peat result in low values of seasonal heat flow and therefore shallow depth of freezing. The maximum depth of frost penetration in the zone of seasonal freezing is much less than for other types of soil and is generally more predictable.

5. In permafrost areas peat exerts an influence out of proportion to the amounts of it that may be present in the soil profile because of its unusual thermal properties. Highly variable moisture contents in surface peat layers at a single site lead to great variations in the rate and depth of freezing and thawing. Field studies have shown that snow cover is equally important in determining seasonal heat flow and depth of freezing.

6. Field studies at Thompson indicate that heat flow meters give useful values for surface heat flow during winter. Other heat balance measurements have also provided useful comparative data. It should be realized by those preparing to make similar studies that the quantities of heat involved and the differences between sites are usually very small and therefore difficult to measure. The fragile nature of peatland also makes this type of terrain very susceptible to disturbance. Daily observations at Thompson inadvertently caused sufficient disturbance at one site to result in thawing of the permafrost, which is very close to 0°C.

Because of the variability in the factors that determine the freezing regime of peatland, normal methods of obtaining site information by closely spaced subsurface investigations may often prove impractical. Under such circumstances engineers require an understanding of the factors that determine freezing and thawing to assess ground conditions. It is hoped that this paper will provide some of this background information on the freezing of peatland.

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TABLE I

CALCULATED HEAT FLOW THROUGH SNOW COVER AND FROZEN LAYERS: 5-26 FEBRUARY 1969, MER BLEUE BOG

Station No.	Relative Elevation (cm)	Snow Depth (cm) 5 Feb	Snow Depth (cm) 26 Feb	Frost Depth (cm) 5 Feb	Frost Depth (cm) 26 Feb	Heat Flow (Q) cal/sq cm
1	+ 23	27	39	22	22	115
2	0	50	55	3	4	99
3	+ 11	40	45	10	11	105
4	+ 25	25	30	22	25	115
5	+ 9	42	47	8	9	103
6	+ 4	45	52	4	7	109
7	+ 20	30	35	24	26	95
8	+ 3	47	47	0	1	103
9	+ 11	37	45	12	14	101
10	+ 16	35	37	16	19	105
11	+ 12	35	40	16	18	102
12	- 1	47	55	3	5	102
13	+ 13	32	37	15	16	115
14	- 5	40	51	5	8	116
15	+ 14	22	31	23	25	120
					Average:	107

TABLE II

FROST PENETRATION, THOMPSON, MANITOBA

	No Permafrost		Permafrost	
	Site A	Site B	Site C	Site D
Rate of Penetration, (cm/week)				
1968 - 69	5.3 (29 Oct - 28 Feb) 3.0 (1 Mar - 20 May)	2.0 (29 Oct - 10 Dec) 4.3 (10 Dec - 25 Feb) 3.0 (25 Feb - 1 Apr)	8.8 (19 Oct - 12 Nov) 1.8 (3 Dec - 15 Jan)	6.8 (15 Oct - 12 Nov) 0.8 (12 Nov - 31 Dec)
1969 - 70	5.5 (10 Nov - 28 Apr) 1.8 (28 Apr - 12 June)	5.0 (17 Nov - 23 Mar) 1.8 (23 Mar - 12 May)	27.0 (20 Oct - 27 Oct) 3.3 (27 Oct - 12 Feb)	14.5 (14 Oct - 27 Oct) 1.3 (27 Oct - 16 Dec)
1970 - 71	5.5 (12 Nov - 26 Feb) 2.5 (26 Feb - 21 May)	4.5 (18 Nov - 26 Feb) 2.0 (26 Feb - 7 May)	11.0 (12 Nov - 23 Jan)	2.5 (18 Nov - 23 Dec) 9.0 (23 Dec - 6 Jan)
Deepest Frost Penetration, cm (Date First Attained)				
1968 - 69	110 (20 May 1969)	68 (15 Apr 1969)	Seasonal frost penetrated to permafrost table at 53-cm depth.	Seasonal frost penetrated to permafrost table at 75-cm depth.
1969 - 70	143 (12 June 1970)	104 (12 June 1970)	Seasonal frost penetrated to permafrost table at 105-cm depth.	Seasonal frost penetrated to permafrost table at 75-cm depth.
1970 - 71	110 (23 July 1971)	80 (7 May 1971)	Seasonal frost penetrated to permafrost table at 123-cm depth.	Seasonal frost penetrated to permafrost table at 75-cm depth.
Beginning of Ground Thawing				
1968 - 69	21 Apr 1969	17 Apr 1969	18 Apr 1969	18 Apr 1969
1969 - 70	9 May 1970	2 May 1970	1 May 1970	28 Apr 1970
1970 - 71	22 Apr 1971	27 Apr 1971	22 Apr 1971	26 Apr 1971
Thawing of Last Frozen Ground				
1968 - 69	23 Aug 1969	18 July 1969	about 15 Jan 1969	about 15 Jan 1969
1969 - 70	12 Sept 1970	2 Aug 1970	14 Feb 1970	31 Dec 1969
1970 - 71	3 Aug 1971	4 July 1971	about 23 Jan 1971	9 Jan 1971

TABLE III

CLIMATIC AND TERRAIN DATA
THOMPSON, MANITOBA

	Site A	Site B	Site C	Site D
Air Temperature, °C				
Winter - mean				
Oct. 1968 - May 1969	-12.1	-12.4	-12.6	-12.3
Oct. 1969 - May 1970	-11.3	-11.4	-11.7	-11.6
Oct. 1970 - May 1971	-11.6	-11.7	-11.9	-11.3
Summer - mean				
June - Sept 1969	13.1	12.7	13.2	12.3
June - Sept 1970	13.6	13.3	13.1	13.0
June - Sept 1971	13.6	13.0	13.4	12.6
Average Annual				
Oct 1968 - Sept 1971	-3.3	-3.6	-3.6	-3.6
Ground Surface Temperature, °C				
Winter - mean				
Oct 1968 - May 1969	-4.3	-4.1	-4.5	-6.4
Oct 1969 - May 1970	-4.9	-6.1	-4.4	-6.6
Oct 1970 - May 1971	-2.5	-4.4	-4.2	-5.8
Summer - mean				
June - Sept 1969	5.8	11.9	9.5	8.8
June - Sept 1970	8.2	11.6	8.9	10.6
June - Sept 1971	9.2	11.9	10.3	10.2
Average Annual				
Oct 1968 - Sept 1971	-0.1	0.7	-1.4	-0.9
Wind Speed, km/hr				
Winter - mean				
Oct 1969 - May 1970	0.40	0.34	0.43	0.50
Oct 1970 - May 1971	0.61	0.37	0.90	0.67
Summer - mean				
June - Sept 1970	0.77	0.51	1.30	0.94
June - Sept 1971	1.31	0.22	0.54	0.77
Ground Heat Flow, cal/cm²/hr				
Winter - mean				
Oct 1969 - May 1970	-0.16	-0.18	-0.23	-0.11
Oct 1970 - May 1971	-0.22	-0.14	-0.13	-0.10
Summer - mean				
June - Sept 1970	0.53	0.48	0.66	0.44
June - Sept 1971	0.53	0.62	0.74	0.58
Surface Organic Layer, (cm)				
Moss/forest litter	10	10	17.5	12.5
Peat	0	10	17.5	0
Organic clay/silt (per cent organic content by weight)	0	23 (31)	10 (26)	17.5 (77)
Total thickness (cm)	10	43	45	30
Snow on Ground				
Thickness - cm - mean				
Oct 1968 - May 1969	29	35	38	21
Oct 1969 - May 1970	28	26	29	23
Oct 1970 - May 1971	39	42	41	33
Density - gm/cc - mean				
Oct 1968 - May 1969	0.198	0.206	0.213	0.186
Oct 1969 - May 1970	0.211	0.197	0.174	0.183
Oct 1970 - May 1971	0.179	0.174	0.180	0.173

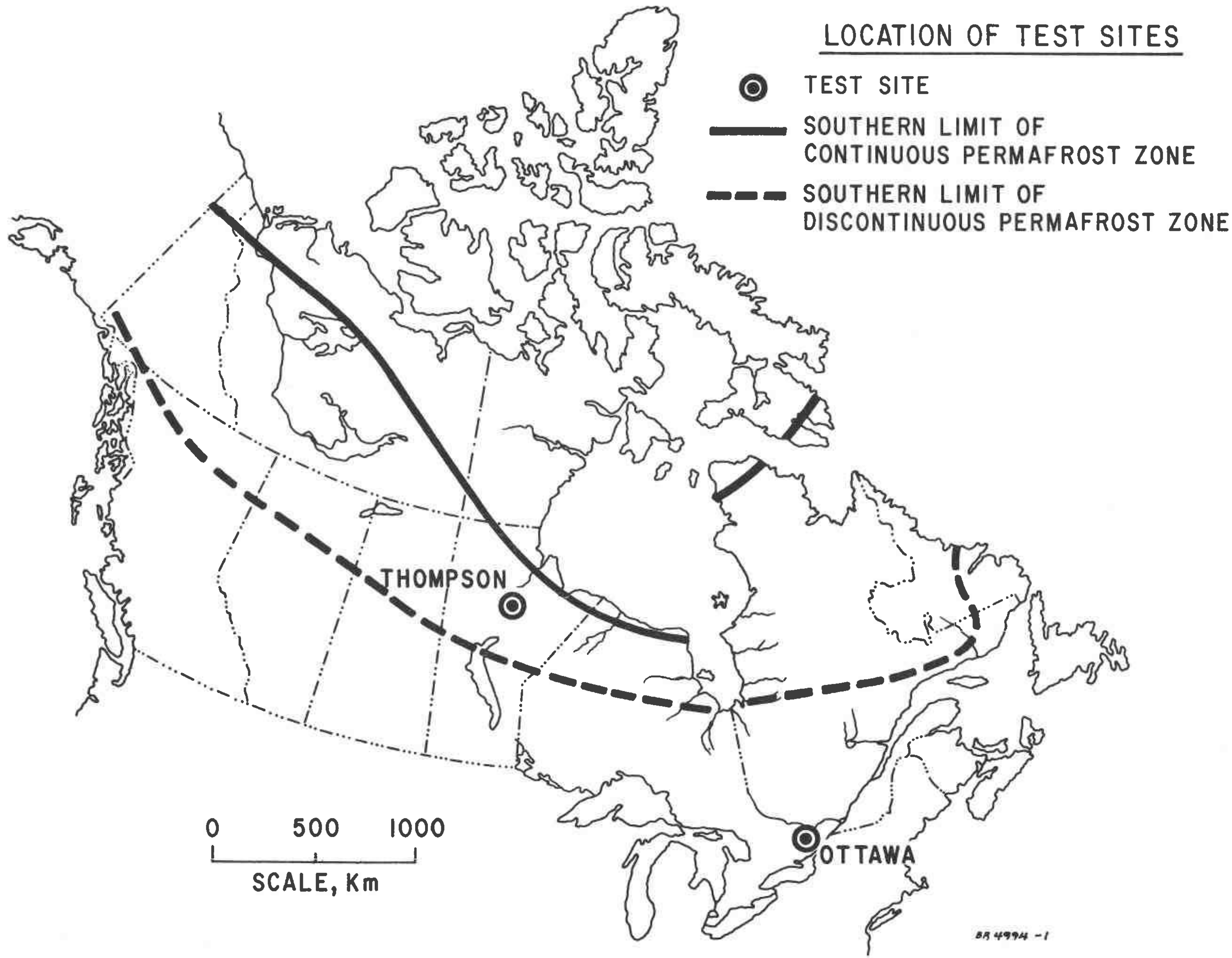
LOCATION OF TEST SITES
TEST SITE
SOUTHERN LIMIT OF CONTINUOUS PERMAFROST ZONE
SOUTHERN LIMIT OF DISCONTINUOUS PERMAFROST ZONE
THOMPSON
OTTAWA
0
500
1000
SCALE, Km
BR 4994 -1

FIGURE 1

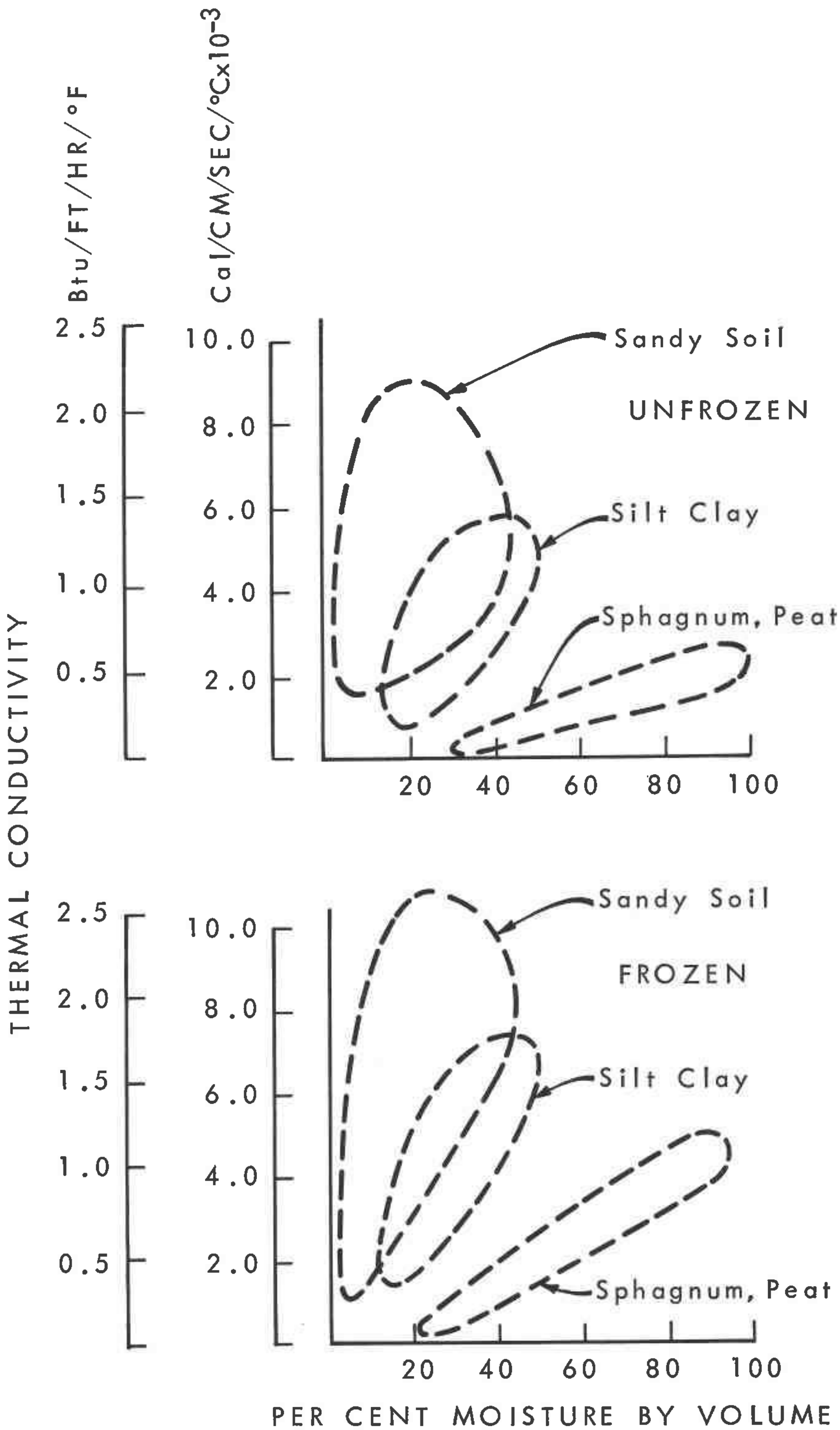

FIGURE 2

VARIATION OF THERMAL CONDUCTIVITY OF SPHAGNUM, PEAT, SILT CLAY AND SAND WITH MOISTURE CONTENT

AB4994-2

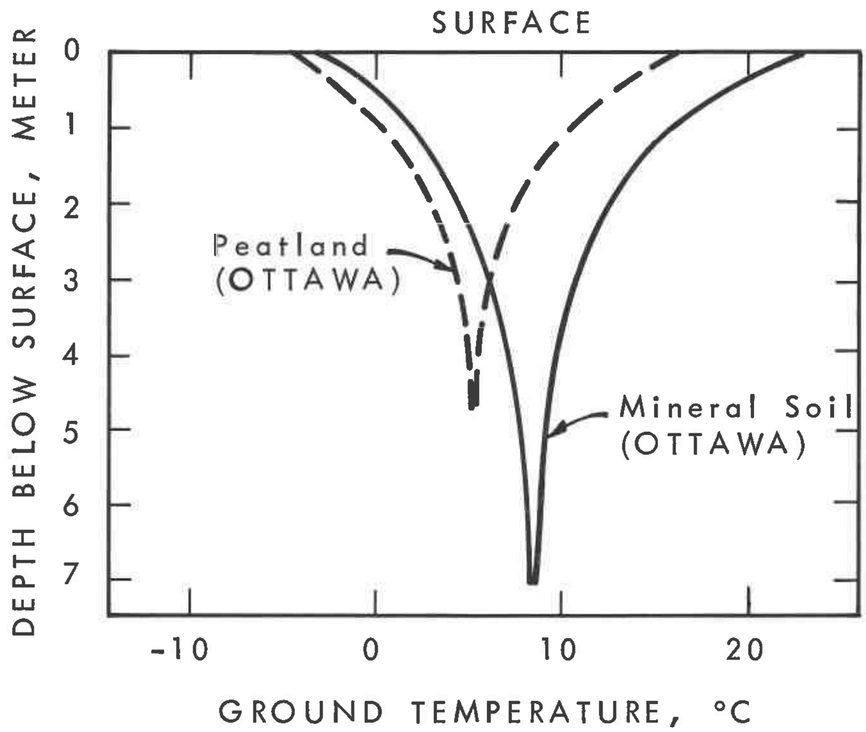

FIGURE 3

COMPARISON OF ANNUAL VARIATION IN GROUND TEMPERATURE-PEAT AND MINERAL SOILS

BR 4994-3

FIGURE 4

Mer Bleue Peat Bog

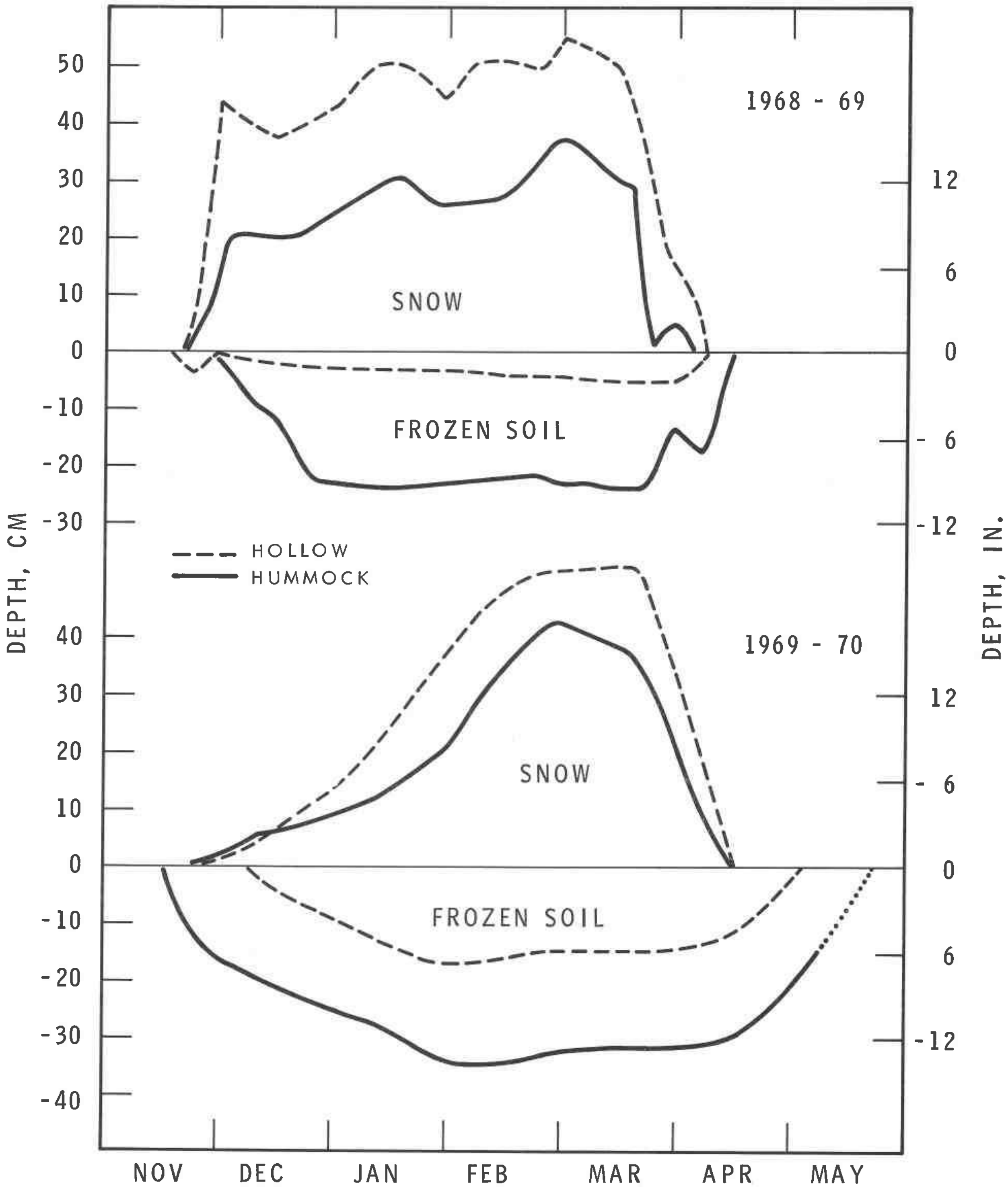

FIGURE 5

SNOW DEPTH AND DEPTH OF FREEZING DURING 1968 - 69, 1969 - 70, MER BLEUE

BR 4909-2

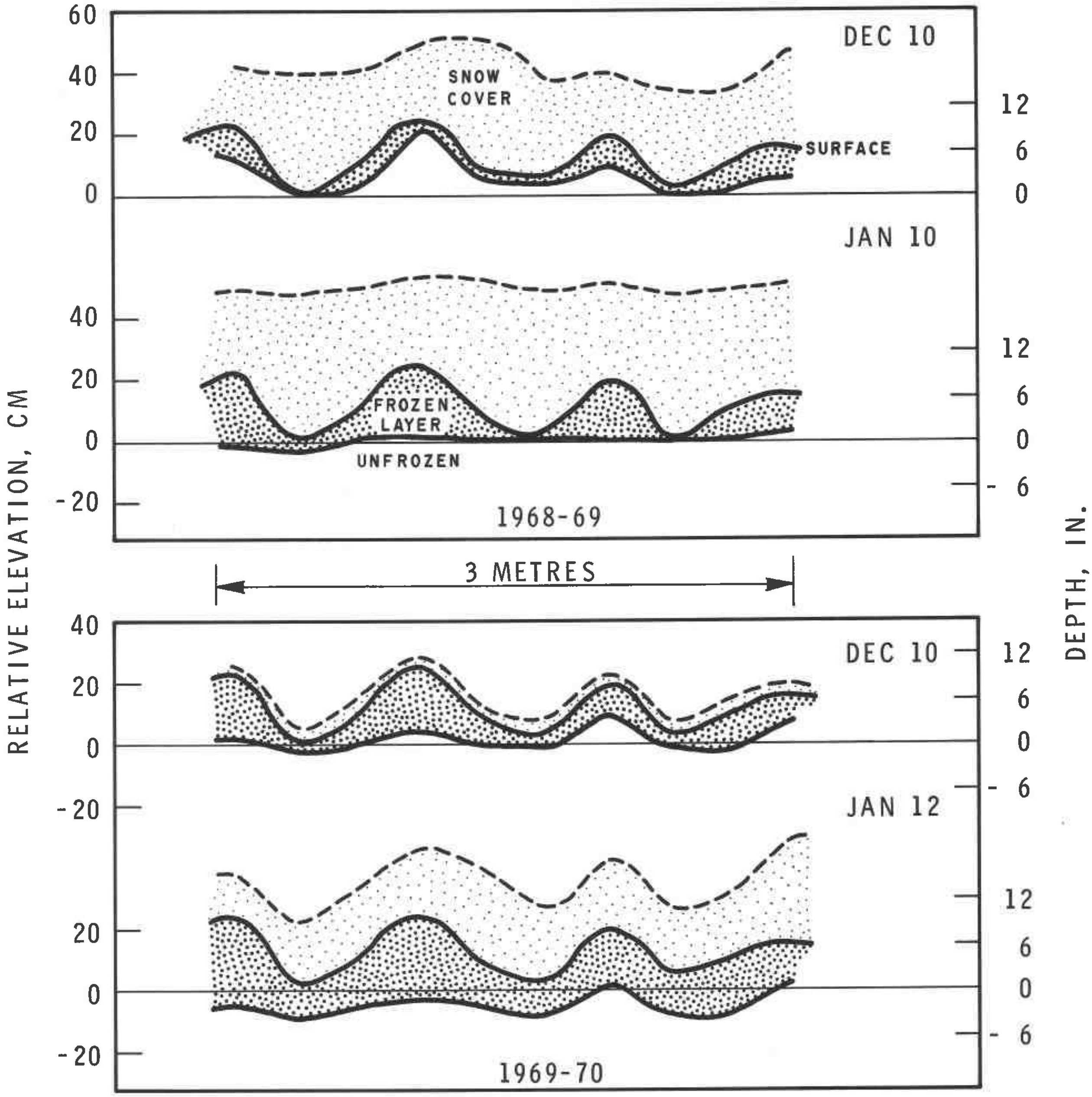

FIGURE 6

SECTION OF FROZEN LAYER SHOWING LOCAL VARIATIONS IN FREEZING PATTERN, MER BLEUE

BR 4909-1

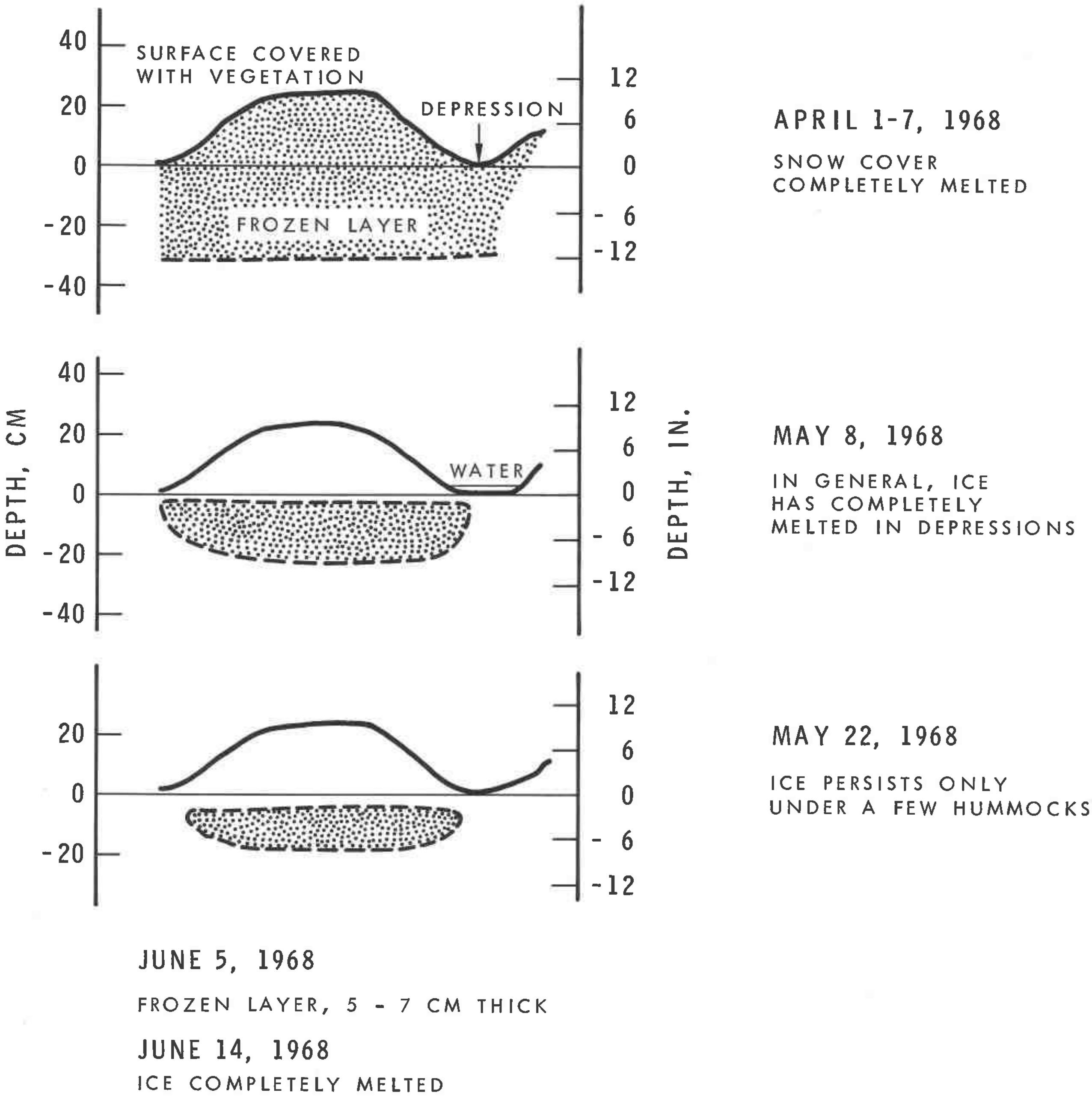

FIGURE 7

TYPICAL THAWING PATTERN, 1968 MER BLEUE

BR 4909 – 3

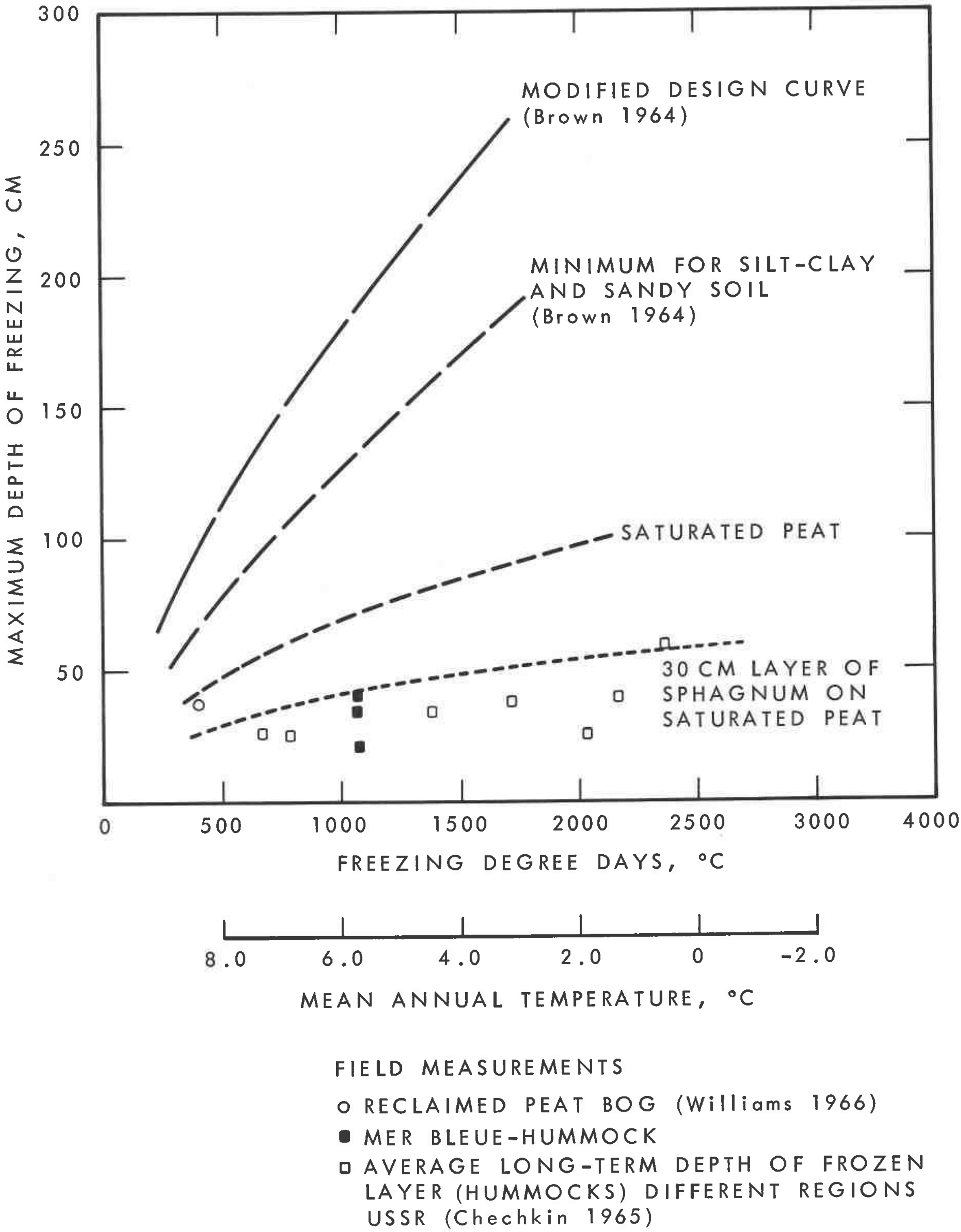

FIGURE 8

COMPARISON OF MAXIMUM DEPTH OF FREEZING

BR4994-4

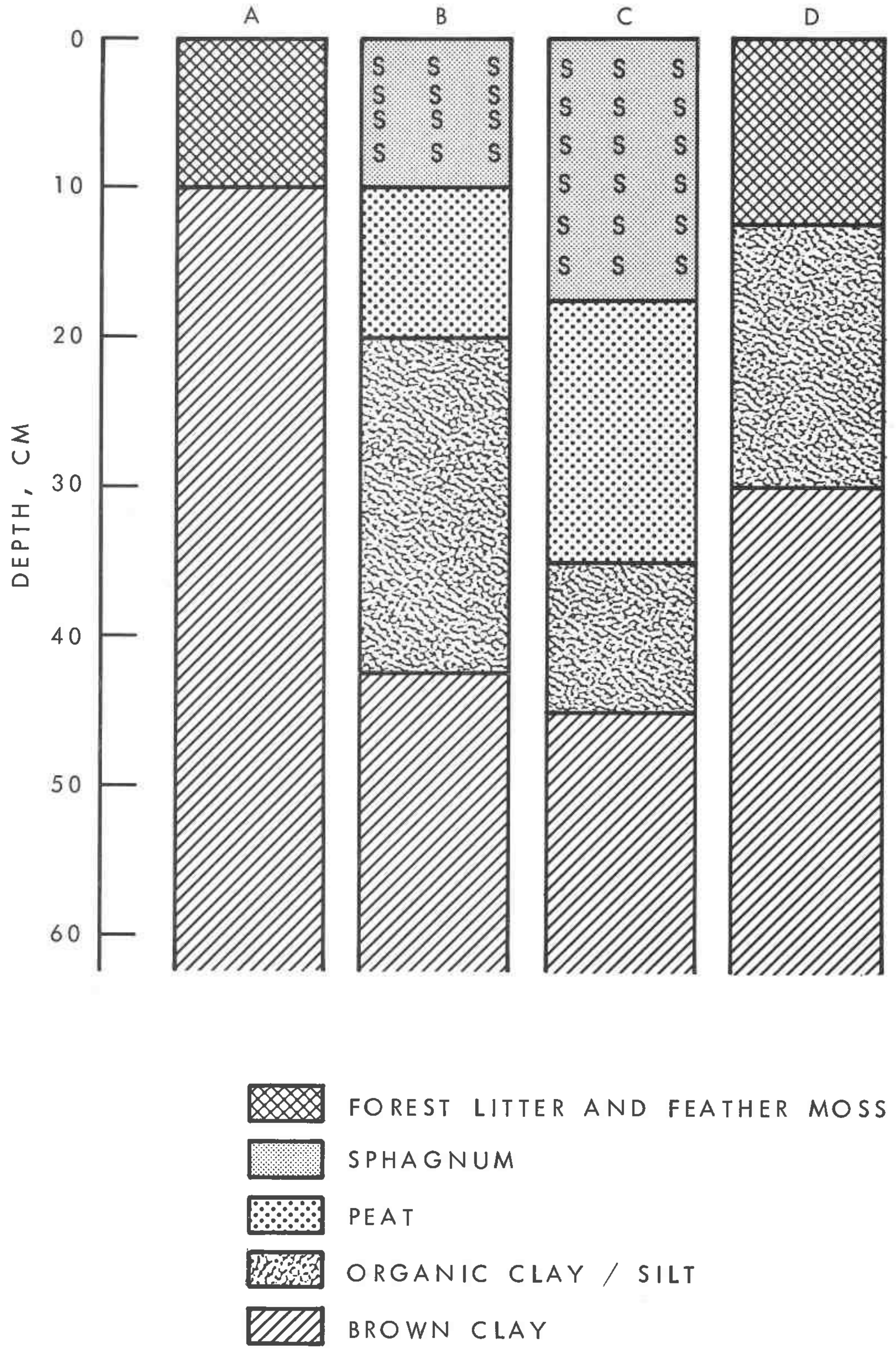

FIGURE 9

SOIL PROFILES OF SITES AT THOMPSON, MANITOBA

BR 4994-5

(a) Site A

(b) Site B

FIGURE 10

Sites With No Permafrost, Thompson, Manitoba

(a) Site C

(b) Site D

FIGURE 11

Sites With Permafrost, Thompson, Manitoba

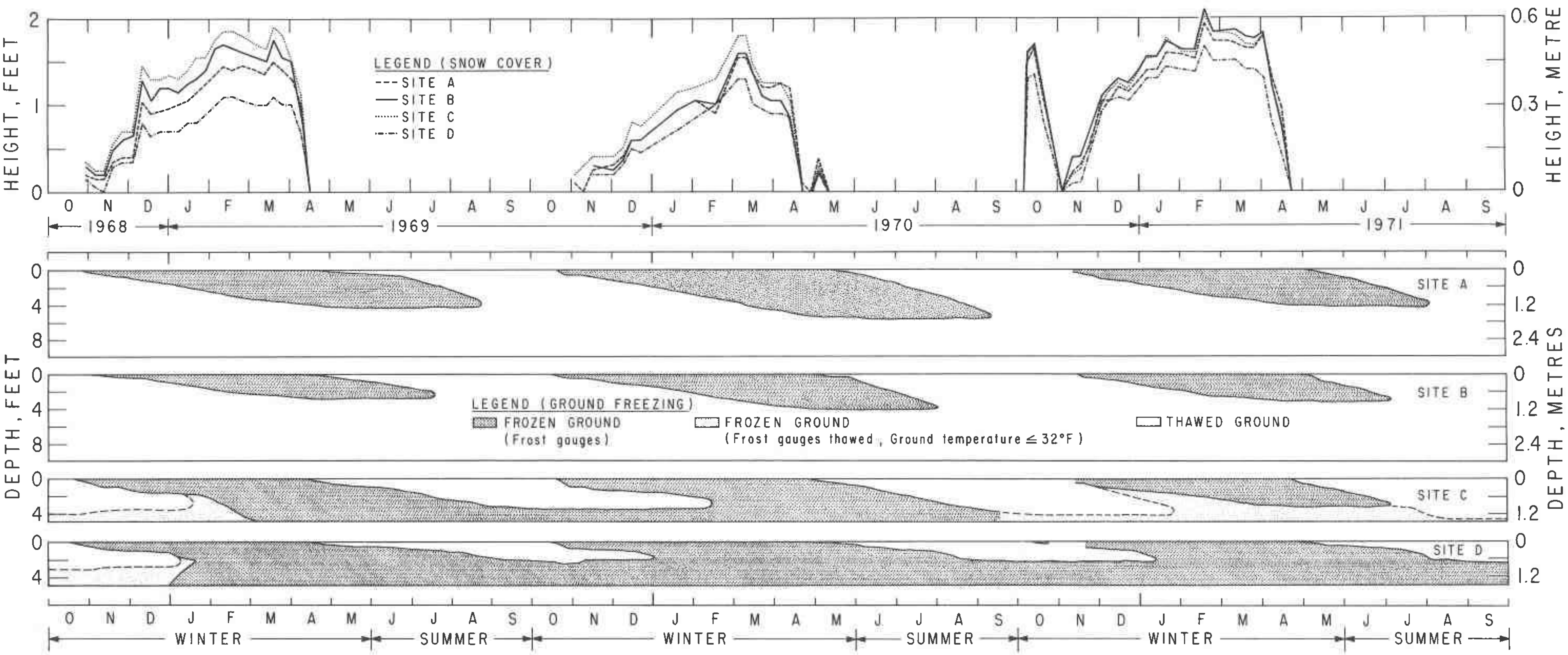

FIGURE 12

SNOW COVER AND GROUND FREEZING REGIMES AT THOMPSON MANITOBA

BR 4909-4

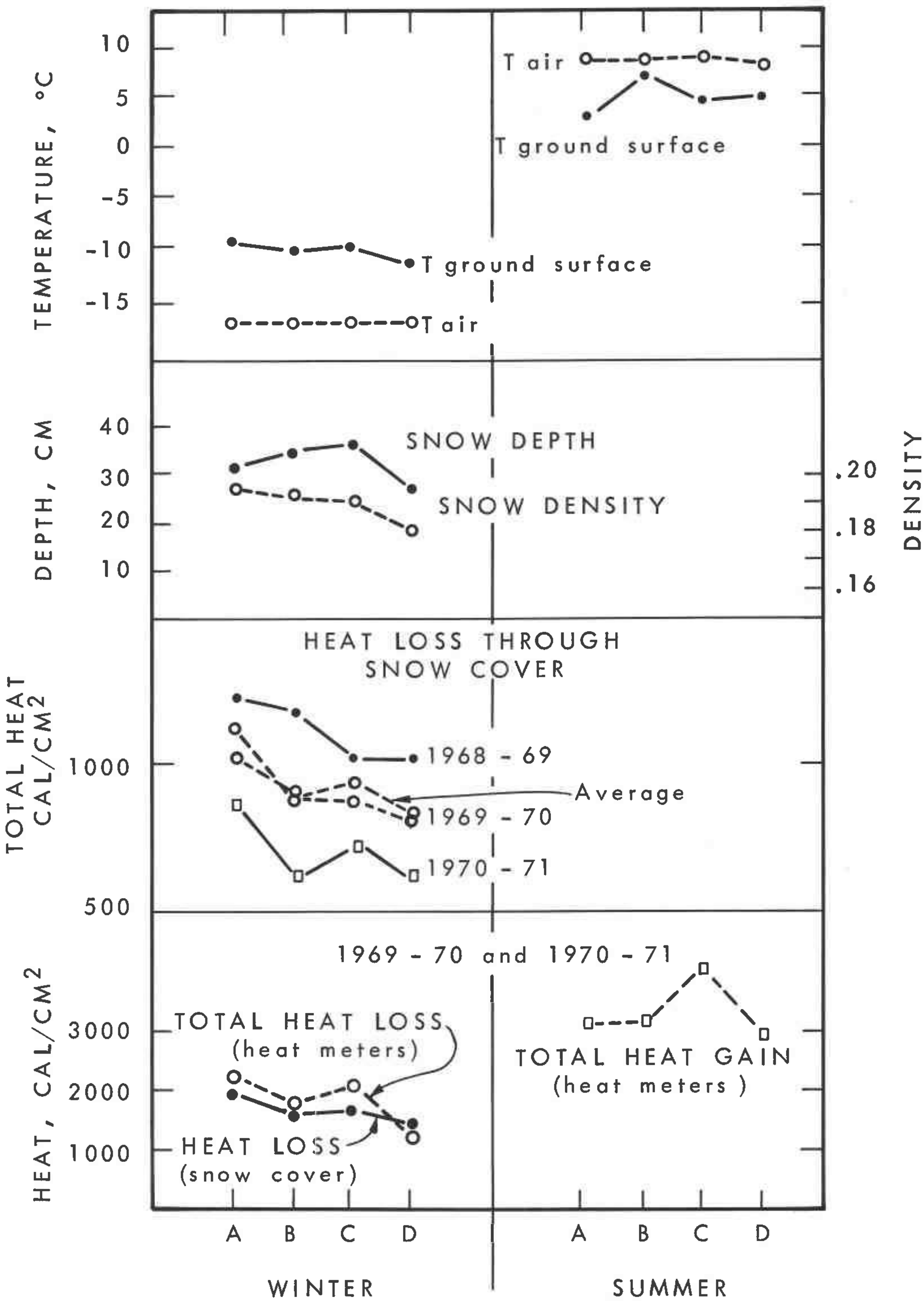

FIGURE 13

SUMMARY OF AIR TEMPERATURE, GROUND TEMPERATURE AND HEAT LOSSES AND GAINS FOR WINTER AND SUMMER PERIODS

BR4994-6