NEAR-SURFACE SOIL HEAT AND WATER FLUXES

IN DIFFERENT PERMAFROST TERRAINS,

CHURCHILL, MANITOBA

© PATRICIA JEAN GOURLAY, B.A.

A Thesis

Submitted to the School of Graduate Studies

in Partial Fulfilment of the Requirements

for the Degree

Master of Science

McMaster University

September, 1981

MASTER OF SCIENCE (1981)
(Geography)

McMASTER UNIVERSITY
Hamilton, Ontario

TITLE: Near-Surface Soil Heat and Water Fluxes in Different Permafrost Terrains, Churchill, Manitoba.

AUTHOR: Patricia Jean Gourlay, B.A. (Wilfrid Laurier University)

ADVISOR: Dr. Wayne R. Rouse

NUMBER OF PAGES: vii, 82.

ABSTRACT

Soil temperatures and active layer depths were recorded near Churchill, Manitoba to determine the near-surface thermal regime for three different permafrost terrains. Soil moisture and net radiation were measured also. The interaction between the heat and water balance of the active layer was investigated to observe the short term effects of water movement on active layer development. Results show that the soil heat component of the surface energy budget composed 18% of the net radiation in the grassland, 14% in the peat and 13% in the tundra. In all terrains at least 93% of the soil heat is consumed in latent heat of fusion. Evaluation of the water budget indicates a large subsurface water loss from the upland tundra and a moderate gain in the grass lowlands during the summer period. Active layer development is shown to respond to both the conductive heat flux and the thermal exports and imports associated with the subsurface water flux.

ACKNOWLEDGEMENTS

This project was the realization of a dream for me and several people must be acknowledged. My sincere thanks are extended to Dr. Wayne Rouse, for giving me the opportunity to go north and for continually offering his encouragement and enthusiasm. Robert Van Eyk was a tremendous help with the field work and I appreciate his assistance during the summer of 1979. The use of the facilities of the Churchill Northern Studies Centre was most helpful and I appreciate the assistance of Bill Erickson, the manager. Dr. John Davies is thanked for his editorial advice and criticisms and for the training his help has provided. To each, I again say thank you.

This study was financially supported by the Department of Indian and Northern Affairs and a research grant from the Natural Science and Engineering Research Council, Canada. Gratefully I acknowledge also the support of an Ontario Graduate Scholarship and a Natural Science and Engineering Research Council post-graduate scholarship.

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

CHAPTER I

INTRODUCTION

1.1 Introduction

Exploration and exploitation of resources in the Canadian north is increasing and is associated with a more concentrated population. Resource-based towns sprinkled across the north will test and stress the precarious ecological balance. A prime determinant of this balance is permafrost, a condition of the ground where the subsurface materials remain below 0 C for more than one year including a full summer season.

Permafrost responds in part to the climate of a particular area; thus depending on the solar energy received and its pattern of usage, the depth and extent of permafrost varies spatially. Permafrost is classified as continuous, discontinuous or sporadic based on the continuity of permafrost. The base of the permafrost is located where the geothermal heat is sufficient to maintain above freezing temperatures. The shallow surface layer which annually thaws and re-freezes is the active layer. The permafrost table describes the top boundary of the permafrost.

The determination of the extent and condition of permafrost, which depends on the terrain type, vegetation cover and aspect is a major endeavour of permafrost research.

Paralleling the growth of permafrost research in the last two decades is the field of energy balance climatology. The fundamental source of energy to the earth is the sun and energy balance climatology investigates the dissipation of the energy through latent, sensible and ground heat transfers. In permafrost regions, the propagation of heat energy through the subsurface is very important since it is used in the thaw process. Hydrological and biological processes and northern vegetation depend on the moisture released after thaw to absorb nutrients from the soil.

Churchill, Manitoba (58^{o} 45' N, 94^{o}04'W) at the confluence of the Churchill River and Hudson Bay, provides a readily accessible site for permafrost research. Located in the Hudson Bay Lowlands, the landscape includes a number of terrain types and microenvironments. This fulfilled the major requirement of a two-pronged research problem. Of initial interest was the examination of the pattern of thaw in different types of terrain as affected by the solar energy input and the mechanisms of ground heat transfer. Soil temperatures during the seasonal thaw period were monitored in three terrain types including a sandy, sparsely vegetated beach ridge designated as the tundra site; the grasslands, a lowland area of wet sedge and the peat site which is found on a palsa. Secondly, the interaction between the heat and water balance of the active layer is investigated to observe the short-term effects of water movement on active layer development.

The research project combines the techniques and theories of energy balance climatology with a problem in permafrost research, which is to comprehend the near-surface thermal energetics in a zone of continuous permafrost. The variation in active layer development in the three terrain types is examined by considering the magnitude of the soil heat term of the energy balance. The effects of the latent heat transfer on the thermal regime helps to explain active layer differences. In addition, the patterns of moisture transfers are studied based on residual calculations from the water budget. Finally, a postulated pattern of soil water transport and the associated heat transfer is offered to explain the terrain-dependent rates and depth of thaw. Identifying the role of heat transfer in the near-surface layers of the terrain contributes to the broad scale understanding of both permafrost dynamics and energy balance climatology in high latitudes.

1.2 Previous research

Scientific interest in the Arctic has experienced a revitalization during the past two decades, responding to the increased demand for natural resources. To minimize the associated ecological damage resulting from terrain disturbances, extensive research programmes began investigating and documenting the natural thermal and vegetational characteristics of the region. A synthesis of the recent research pertinent to this study involves three areas of study, including early descriptive works, initial energy balance investigations and original attempts to simulate the thermal regimes.

The pioneering investigation into seasonal changes in the temperature gradients and active layer depths at Fort Churchill by Brown Beckel (1957) concluded that a deeper thaw occurred with thinner peat layers. In addition, it was found that the removal of plant cover increased the depth of the active layer. This study included 30 sites in both wet and dry environments and provided an early example of comparative terrain studies. The intensive economic development in the Mackenzie Valley prompted research on the thermal characteristics of the soils including thermal conductivity and heat capacity (Judge, 1973). Brown (1978) reported on a long term project investigating soil temperatures along a transect through northern Manitoba and the Keewatin District, N.W.T. The goal was to discern the role of climate and terrain on regulating soil temperatures and thaw depths. These "data-gathering" projects served to illuminate needy research areas, one of which was the behaviour of the surface energy budget in permafrost terrain.

The preliminary northern energy balance research originated at Barrow, Alaska. Kelley and Weaver (1969) investigated the temperature stratification in the snow and subsurface as it related to net radiation. Weller and Holmgren (1974) improved on this work at Barrow when they determined the seasonal variations in the magnitude of the component fluxes of the surface energy budget. Recent studies addressed the effect of surface vegetation and snowcover on permafrost distribution. Results confirm that the disturbance or removal of surface vegetation leads to increased depths of thaw resulting from a decrease in the latent heat component of the substratum (Addison and Bliss, 1980; Haag and Bliss, 1974; Thompson and Fahey, 1977). The knowledge of permafrost distribution in Canada was advanced by the series of papers presented in a volume of Geographie Physique et Quaternaire (1979) which was devoted to the permafrost of Quebec-Labrador. Contemporary permafrost is found to exist as far south as the Chic-Chocs mountains of the Gaspé (Gray and Brown, 1979).

More practical applications of the energy balance technique include projects of permafrost amelioration through the alteration of the natural surface heat balance (Nicholson, 1976). Nicholson achieved limited success and subsequently permafrost amelioration through surface alterations has proved feasible for small-scale projects.

Mathematical modelling of the near-surface thermal regimes in the arctic began in the early 1970's. One-dimensional models of heat conduction based on assumptions of soil homogeneity and no mass transfer (Van Wijk and de Vries, 1963) were improved upon by Nakano and Brown (1972) by including a "latent heat function" to handle the phase change of water in permafrost. This extra term incorporated the heat capacity of the soils with the latent heat function for a known ice content. They concluded that the latent heat function of organics is especially important to the thermal regime.

Noted success in simulating ground thermal regimes has been achieved by Smith (1975 a,b) based on the earlier work of Outcalt (1975, 1972). Smith's model uses standard meteorological parameters and site specific information such as the volume fraction of organics, minerals or water in each soil layer. Snowmelt rates, active layer depths and freeze-back rates may be accurately predicted. Smith notes the practical application of such a model for land-use planning problems in the north.

Subsequent models concentrate on the coupled nature of the thermal and moisture transports in the arctic soils (Guymon and Luthin, 1974; Luthin and Guymon, 1974). The importance of vegetation cover and surface organics as modifiers of the thermal and moisture conditions is discussed.

Ng and Miller (1977) elaborate on these relationships in their work on vegetation and soil temperatures. They suggest further work must concentrate on the thermal conductivity of the organic layer and the resistance of the subsurface to evaporation.

Several researchers presented the results of their modelling efforts at the Third International Conference on Permafrost in 1978. Based on data from wet tundra soils near Barrow, Alaska, McGaw, Outcalt and Ng (1978) report that thermal conductivity is the prime determinant of summer heat flow into the active layer. A model for more practical applications is presented by Abbey _et al_ (1978) who empirically determined a model with readily measured parameters to predict ground heat flux. It is based on cumulative net radiation and cumulative air temperatures.

The last decade witnessed an intense concern for determining quantitatively the thermal regime in northern terrains. In their review of permafrost research, Gold and Lachenbruch (1973) point out the need for further research of this type by suggesting that "investigations should be undertaken to establish the relationship between the characteristics of the surface and near-surface materials and the components of the surface heat exchange." At the Churchill site, research on these problems continues.

CHAPTER II
Theoretical Background

2.1 Energy Budget

The thermal regime of a site depends on the external energy receipt and the internal composition of the subsurface material. In areas of permafrost, the cold lower boundary of the active layer maintains cool soil temperatures in contrast to the surface where temperatures may exceed 30 C. These conditions produce an unique summer time thermal regime.

The major energy source is solar radiation, and depending on latitude, season and atmospheric composition the radiation at the surface is less than that received at the top of the earth's atmosphere. The surface radiation balance, or net radiation, is given by

$$Q^* = K\downarrow (1-a) + L\downarrow - L\uparrow \qquad (1)$$

where Q^* = net all-wave radiation
$K\downarrow$ = total incoming solar radiation
$L\downarrow$ = total incoming longwave radiation
$L\uparrow$ = emitted longwave radiation
a = surface albedo

Surface albedo and emitted longwave radiation are site specific. Both are controlled by surface cover and moisture content. The albedo value of natural vegetated surfaces of the Churchill area including a wetland, lichen heath, and a wood/lichen site has an average of 0.12. Petzold and Rencz (1975) calculated albedo for several subarctic surfaces finding an albedo of 0.21 for a heath lichen tundra, and 0.11 for a sedge-moss bog.

No reference was made to a peat dominated surface. The second term, L↑, depends on surface temperature, such that

$$L\uparrow = \varepsilon\sigma T_s^4 \qquad (2)$$

where ε = emissivity
σ = Stefan-Boltzmann constant
T_s = surface temperature

At the atmosphere-surface interface, the net radiation must be dissipated according to the laws of energy conservation. The transfer and transformations of this energy may be described by

$$Q^* - Q_e - Q_h - Q_g = 0 \qquad (3)$$

in which Q^* is the source of energy which fuels the latent (Q_e), sensible (Q_h), and soil heat (Q_g) fluxes normally transferring energy away from the surface when Q^* is positive. Other processes such as photosynthesis are of a much smaller magnitude and are often neglected. Here, the non-radiative fluxes are considered positive when energy is directed away from the surface, thus a positive soil heat term indicates a gain of heat energy in the substrata. In areas of permafrost, Q_g is the main source of energy that controls the freezing and thawing of the active layer. The magnitude and direction of the fluxes may vary during the year with all fluxes commonly being negative in the long arctic winter. In contrast, the high net radiation in the summer is due to long daylight hours and small albedos.

The tundra microclimate at Barrow, Alaska studied by Weller and Holmgren (1974) experiences distinct patterns of energy consumption throughout the year. Spring snowmelt produces the most dramatic changes in the energy balance with net radiation increasing significantly at this time. During snowmelt they found that 9% of the Q^* was used to heat the soil, compared to only 2% at midsummer. Energy expended in evaporation experienced similar fluctuations varying from 73% of Q* immediately following melt to 66% at midsummer. This initial investigation of the energy balance fluctuations of an arctic tundra provided the basic data for further studies.

The transfer of heat energy into the subsurface is expressed as

$$Q_g = \lambda \frac{dT}{dz} \qquad (4)$$

where dT/dz is the temperature gradient, and λ is the thermal conductivity of the soil material. In a conclusive laboratory study, Kersten (1949) showed the dependence of thermal conductivity on soil composition. In the varying terrains of the Churchill region, a wide range of thermal conductivity values is expected.

The storage of heat energy in the ground, in the absence of any advective effects, depends on the composition and moisture content. The heat energy conducted into the ground will produce a temperature change according to the heat capacity, C, of the material. For a soil layer, change in heat storage is

$$\frac{dS}{dt} = \int_{z_1}^{-z_2} \frac{\partial (CT)}{\partial t} \, dz \qquad (5)$$

where $\frac{dS}{dt}$ = change in stored soil heat energy over time period dt

C = $C\ (t,z\ )$ heat capacity

$\frac{\partial T}{\partial t}$ = rate of temperature change over time

dz = depth increment

In order to conduct the measurement programme, the right-hand side of eq. (5) was approximated by

$$\frac{dS}{dt} = \sum_{i=1}^{n} C_i \left(\overline{\frac{\Delta T}{\Delta t}}\right)_i \Delta z_i \qquad (6)$$

where n is the number of soil layers used in the computations, and n varies over the season.

To determine the heat storage in an entire soil profile, the energy leaving the bottom layer must also be considered. In permafrost regions, the bottom boundary of the soil profile in the active layer is maintained at a temperature of 0 C. During the late summer when soil heat storage is reaching its maximum, the temperature gradients at this level are weak (Figs. 2.1) and the heat flux is assumed to be zero. Heat transfer in dry, unfrozen, vegetated soils is almost solely due to conduction; however in wet and frozen soils, heat is transferred by water movement and through the latent heat of phase changes. Hence, a more complete calculation of heat storage must consist of at least five energy terms including: the surface heat flux, the heat flux across the frost table, the latent heat of fusion consumed in the thawing of ground ice, the downward heat transfer of percolating rainwater and an advective component of soil heat.

This advective term may be conceptualized as including lateral heat transfer by conduction and by horizontal moisture movement. In many cases, the horizontal movement of water exceeds the rainfall input and this advective term becomes a negative quantity indicating the loss of heat energy through water movement in the subsurface. Evidence to support this concept is discussed in section 5.2.

Different soils are characterized by their heat capacities. Following de Vries (1963) heat capacity may be expressed as a linear function of volume fractions and heat capacity of the soil constituents such that

$$C = X_m C_m + X_o C_o + X_w C_w \tag{7}$$

where X_m, X_o, X_w are volume fractions of minerals, organics and water

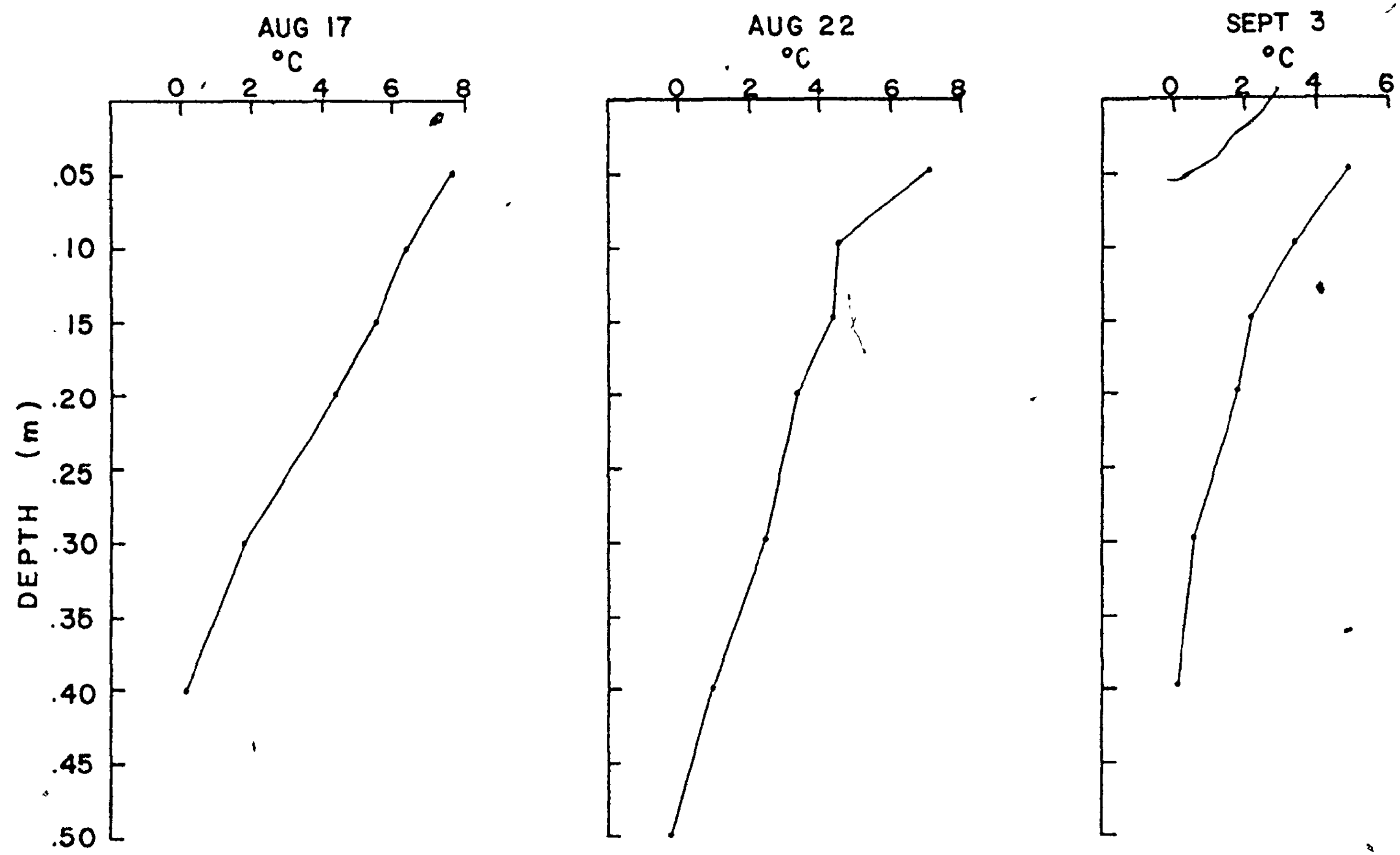


FIG 2.1(a) SAMPLE LATE SUMMER TEMPERATURE PROFILES - PEAT

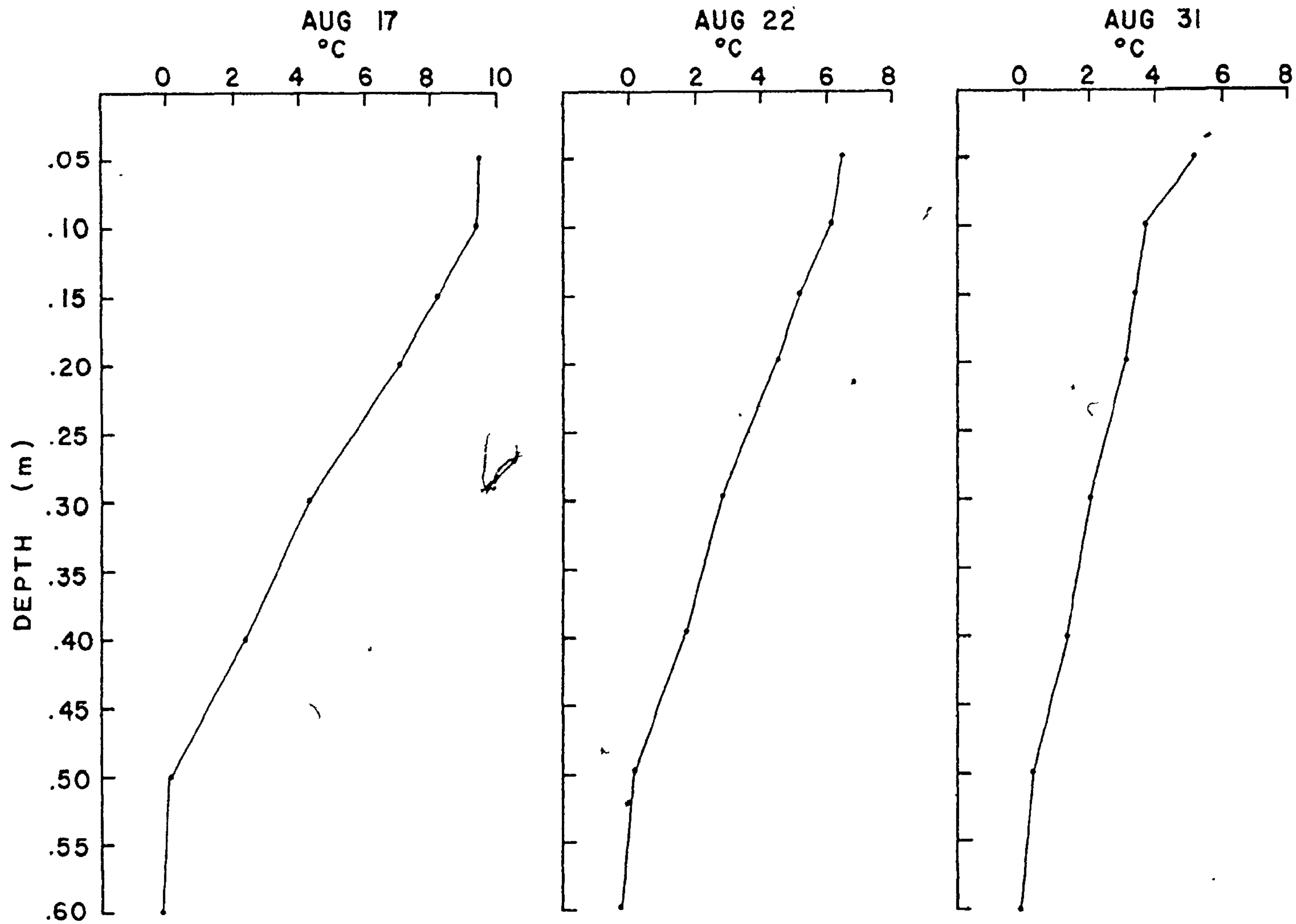


FIG 2.1(b) SAMPLE LATE SUMMER TEMPERATURE PROFILES - GRASSLAND

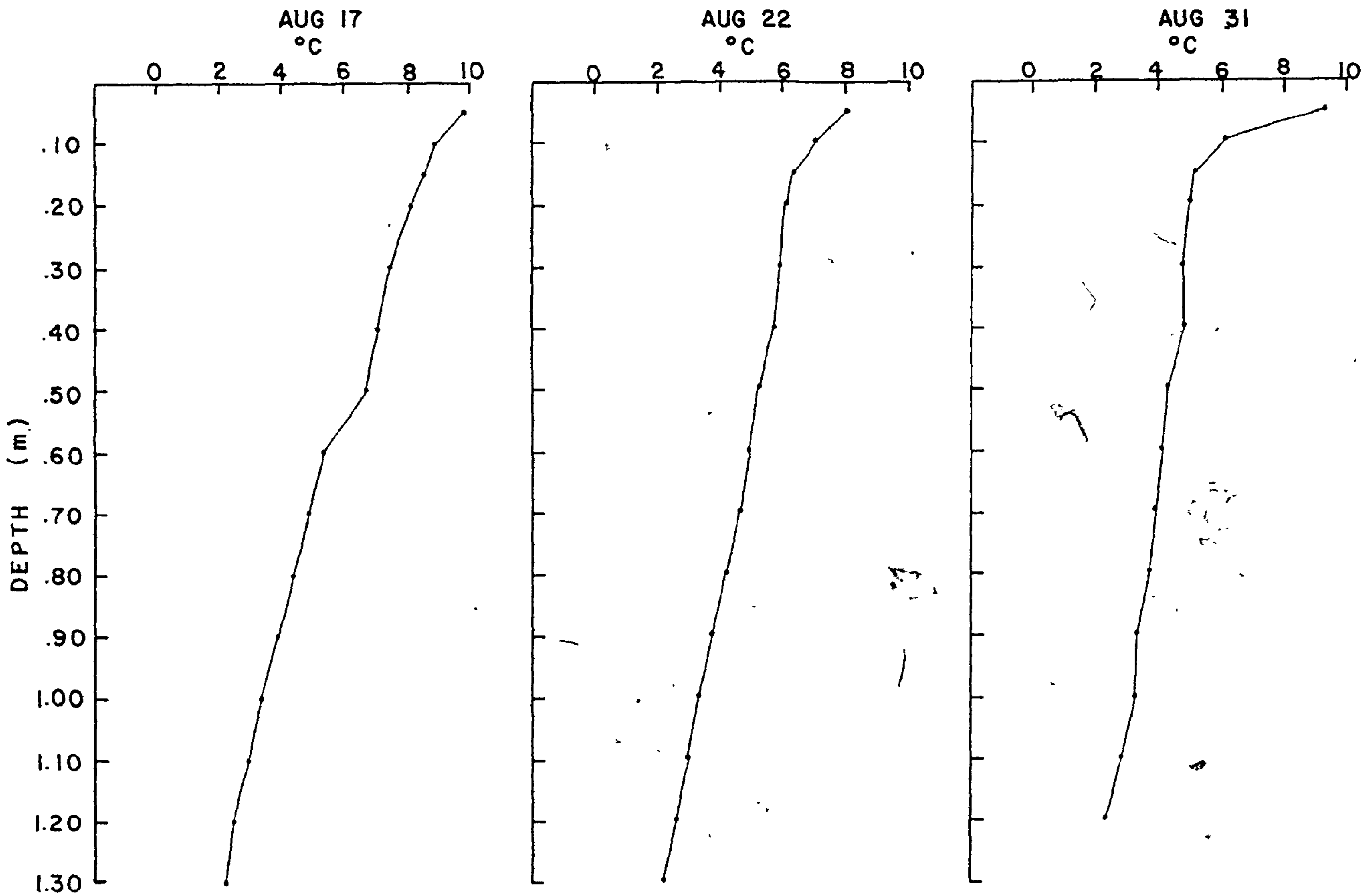


FIG 2.1(c) SAMPLE LATE SUMMER TEMPERATURE PROFILES TUNDRA

respectively. De Vries (1963) found on average that

$$\begin{aligned} C_m &= 1.93 \times 10^6 \ Jm^{-3}K^{-1} \\ C_o &= 2.51 \times 10^6 \ Jm^{-3}K^{-1} \\ C_w &= 4.18 \times 10^6 \ Jm^{-3}K^{-1} \end{aligned} \tag{8}$$

In permafrost where ice is encountered, the heat capacity decreases abruptly since the heat capacity of ice is less than one-half the heat capacity of water, $C_i = 1.92 \times 10^6 \ Jm^{-3} K^{-1}$. Therefore, in heat capacity calculations including ice, XiCi was exchanged for XwCw. The final equation became

a) for thawed soils:

$$C = 1.93 \times 10^6 \ Xm + 2.51 \times 10^6 \ Xo + 4.18 \times 10^6 Xw \tag{9}$$

b) for frozen soils:

$$C = 1.93 \times 10^6 Xm + 2.51 \times 10^6 Xo + 1.92 \times 10^6 Xi \tag{10}$$

Soil moisture plays an important role in buffering the thermal regime of a soil. The sensitivity of the soil temperatures to increased amounts of absorbed heat energy diminishes at high moisture contents. In wet soils the high heat capacity permits only a modest temperature change per unit of heat input. Similarly, with increased soil moisture, the thermal conductivity will increase as water supplants air in the pore space. Optimum heat exchange occurs with 8% to 20% moisture content. Above this amount, increased moisture results in decreasing thermal diffusivity, $K = \lambda/C$. In Table I, standard values of these soil parameters are presented. For comparison purposes, sand with 40% pore space and peat with 20% pore space are most similar to the ~~soils~~ under investigation.

The remaining terms of the energy budget are the latent and sensible heat fluxes. High latitude evaporation studies (Rouse and

Material	Heat Capacity ($Jm^{-3}K^{-1} \times 10^6$)	Thermal Conductivity ($Wm^{-1}K^{-1}$)	Thermal Diffusivity ($m^2s^{-1} \times 10^6$)
Sandy soil dry	1.28	0.30	0.24
40% pore space saturated	2.96	2.20	0.74
Peat soil dry	0.58	0.06	0.10
80% pore space saturated	4.02	0.50	0.12
Ice	1.93	2.24	1.16
Water	4.18	0.57	0.14
Air	0.0012	0.025	20.50

TABLE I Thermal Properties of Natural Materials (adapted from Oke (1978))

Stewart, 1972; Stewart and Rouse, 1976) detail how evaporation may be evaluated successfully using a modified form of the equilibrium evaporation model. The form of the model is

$$Q_e = \alpha \, (S/(S + \gamma))(Q^* - Q_g) \qquad (11)$$

where S is the slope of the saturation vapour pressure curve at mean air temperature and γ is the psychrometric constant. The empirical parameter α, dependent on both atmospheric and surface conditions, relates actual to equilibrium evaporation.

The extent and nature of the surface vegetation exerts a strong control on evaporation. Unlike temperate regions where soil moisture is a limiting factor, evaporation in the north is inhibited to a greater degree by the xeric vegetation (Rouse and Kershaw, 1973). Lichens do not transpire in contrast to vascular plants. Thus on terrains where lichens are present, less-than-potential evaporation occurs. Rouse (1981b) has determined recently that for tundra at Churchill where evaporation is restricted by a xerophytic surface cover, $\alpha = 0.94$ which closely agrees with earlier high latitude evaporation studies. This value has been applied to both the tundra and peat sites where the covering of lichens and mosses restricts evaporation. In cases where evaporation is uninhibited, Stewart and Rouse (1976) substantiated the use of an $\alpha = 1.26$ in the Q_e calculation. This value was applied to the grassland site.

After determining the latent and ground heat fluxes, the sensible heat flux, Q_h is derived as a residual. With all terms of the surface energy balance computed, a similar budgetting technique is used to determine the magnitude of the water balance components.

2.2 Water Budget.

The temperature regime is determined for the most part by heat conduction but considerable amounts of energy may be transported by water movement. This is especially true in the active layer where steep hydraulic and temperature gradients persist. Guymon (1976,1975) examined the interaction between pore-water pressure and soil temperatures in a lowland and tundra site in Alaska. He noted that "the downward movement of water in the soil profile may convect appreciable quantities of heat, further causing the active layer to melt" (Guymon, 1975). Even in frozen soils of the winter Guymon (1976) suggests that due to the escape of moisture from the soil surface and the development of ice lenses, the resultant low pore-water pressures near the surface maintain steep upward hydraulic gradients. These soil moisture fluctuations must be considered in evaluating the thermal regime.

The water balance of a soil block is in constant equilibrium, balancing inputs against losses from the block. This relationship is expressed as

$$\Delta Sm = P - E + \Delta W_z \qquad (12)$$

such that the change in soil moisture storage ΔSm, equals the precipitation P, minus the evaporation E, and the loss due to horizontal subsurface flow at depth, ΔW_z. Since the subsurface flow is to be calculated as a residual commodity, eq.(12) may be rearranged such that

$$\Delta W_z = \Delta Sm - P + E \qquad (13)$$

where $\Delta Sm = Sm(t_2) - Sm(t_1)$ in which t_1 and t_2 denote initial and subsequent time periods of measurement.

The effect of subsurface water transport on the thermal regime of different terrain units is ascertained by following these procedures. These unique thermal conditions result in variable rates and depths of thaw of the active layer.

CHAPTER III
Site Description and Research Methodology

3.1 Climate

Hudson Bay strongly influences the weather and climate of the Churchill locale. Its presence maintains the arctic-like conditions of Churchill despite its relatively southern location.

Hudson Bay remains ice-covered for the greater part of the year; shorefast ice may last until late June in the Churchill area. The counter-clockwise circulation of the water often drags ice from the north southward during July (Jordan, 1980). Fog is a persistent phenomenon as cold air currents blow across the ice-covered bay. Hudson Bay acts as a large ice-wedge, splitting the continental landmass, thus disrupting the prevailing westerly airflow and introducing a strong northerly airflow. Consequently, Churchill experiences winds which, if they blow from WNW to NNE, originate over cold water. Frequent weak storm cells may develop over the Bay contributing to the unsettled weather. Wind shifts do occur occasionally, causing incursions of warm southern air. Mean daily air temperatures may exceed 20C and reach maxima of 33C as experienced for a four day period in July, 1979.

Continuous climate records have been maintained for Churchill

	MAY	JUNE	JULY	AUGUST	YEAR
<u>SUMMER OF 1979 (°C)</u>					
Mean Daily Maximum	1.8	11.8	17.0	13.3	
Mean Daily Minimum	-5.7	1.4	7.6	4.9	
Mean Daily Temperature	-2.0	6.6	12.3	9.1	
<u>LONG TERM RECORDS</u> (°C)					
Mean Daily Maximum	1.1	10.6	16.9	15.5	-3.3
Mean Daily Minimum	-5.8	1.6	6.9	7.4	-11.3
Mean Daily Temperature	-2.3	6.1	12.0	11.5	-7.2
<u>PRECIPITATION</u> (mm) 1979		52.5	40.3	54.9	
Long Term		40.1	49.0	57.7	

TABLE II. Monthly Climate Summary for Churchill, Manitoba. (Hare and Thomas, 1979; Environment Canada Weather Observations, Churchill "A" Station, 1979).

since 1943. These data, summarized in Table II, were recorded at the municipal airport, 7 km west of the research site. Compared to these long term averages, the weather during the study period was normal with respect to temperature and rainfall. On a monthly basis, the June mean daily temperature of 6.6 C was only slightly warmer than the 6.1 C average. Similarly, July was only 0.3C warmer than normal but on several occasions extreme high maxima of 29-34C were attained. August was cooler than average and three new record lows were noted. July and August were both drier than normal, but the precipitation in June exceeded the longer term norm by 12.4 mm. As temperature and precipitation patterns were normal, the thermal and moisture regimes described here may also be considered representative for the region.

3.2 Geology

The bedrock of the Churchill area consists of Precambrian quartzite ridges varying from 8 to 31 m in height which extend parallel to the shoreline (Brown Beckel, 1957). Inland from the coast spreads a lowland area of marine clays, remnants from the post-glacial invasion of the Tyrell Sea. Other evidence of recent glaciation is documented in rock striae and non-sorted debris. The Laurentide ice sheet retreated only 7-8000 years B.P. from this region (Parsons, 1980).

At present, numerous relic beach ridges dissect a flat plateau. Rapid isostatic uplift at a rate of approximately 1m per century has gradually exposed newer ridges. Drainage is poorly developed contributing to the 75% covering of sedge wetlands and lakes in the area (Brown Beckel, 1957).

The southern boundary of the continuous permafrost zone passes to the south of Churchill as governed by the mean annual air temperature of -7.2C and a mean annual ground temperature of -2.5C (Hare and Thomas, 1979; Brown, 1978). The depth of the active layer is spatially variable in response to the local terrain.

In areas where thick peat insulates the surface, frozen ground may exist at shallow depths of 0.30 m. In contrast, drill hole evidence indicates that beneath bedrock the active layer may exceed 6 m (Parsons, 1980). Although disputed by French and Gilbert (1981) who speculate that the permafrost in the Churchill area may be discontinuous, at the research site, permafrost is encountered in all locations at a depth less than 2 m.

3.3 Research site

The research site, located 10 km east of the Churchill town site and 2 km south of the Hudson Bay coastline, extends over a triangular area of 2400 sq. m (Fig. 3.1). To the south of one of the numerous relic beach ridges lies a lowland region of small ponds and poorly drained grasslands, interspersed with small peat palsas. The individual research sites were chosen randomly while the area was snowcovered. Two separate study sites of each terrain type were monitored and the measurements were averaged. The terrain units discussed are referred to as the peat, grassland and tundra sites (Fig. 3.2).

The most common terrain type is the grass lowland. As shown in Fig. 3.3 (a), the grasslands are poorly drained due to the flatness and the underlying permafrost. As a result, these soils remain at or

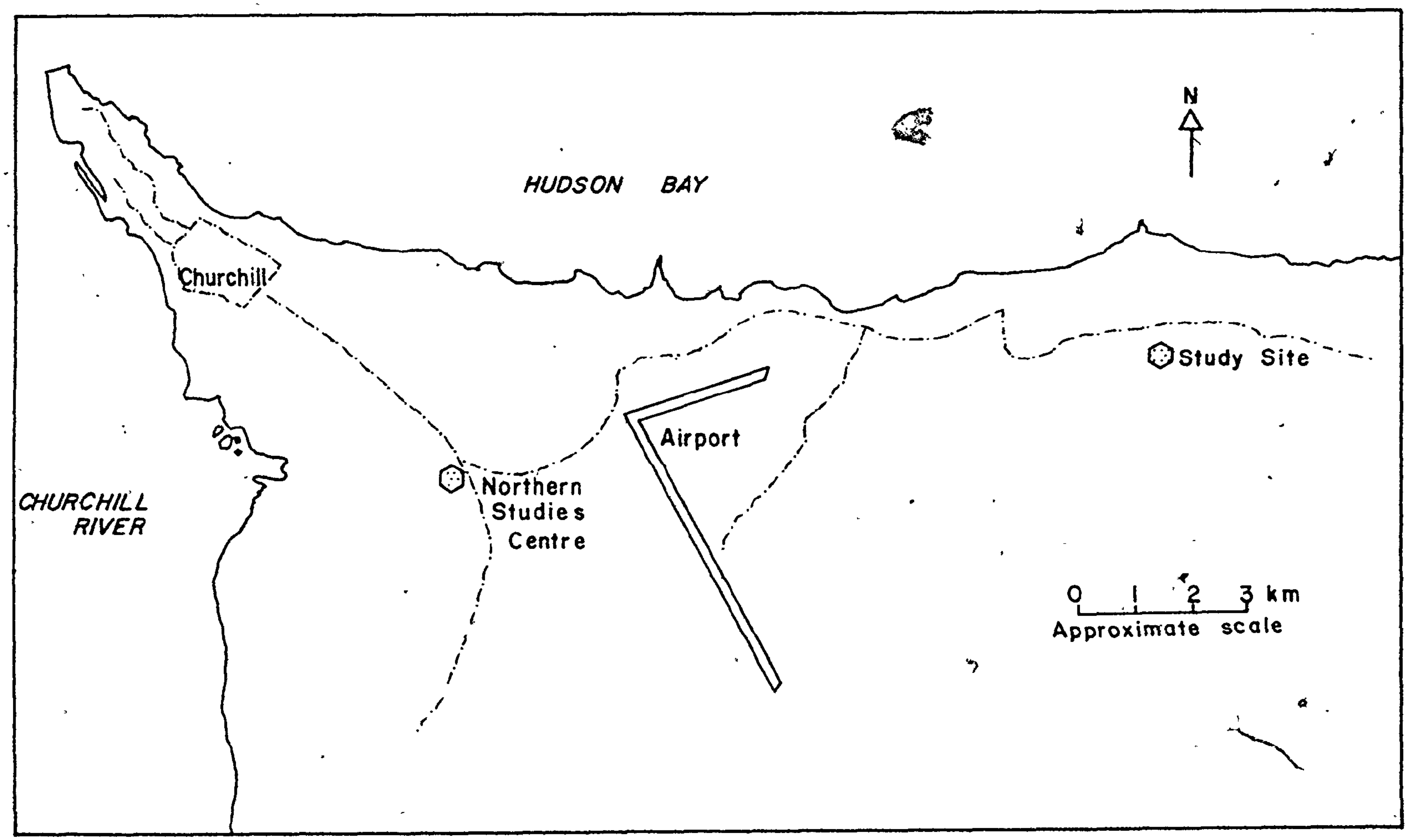


FIG 3.1 STUDY SITE LOCATION

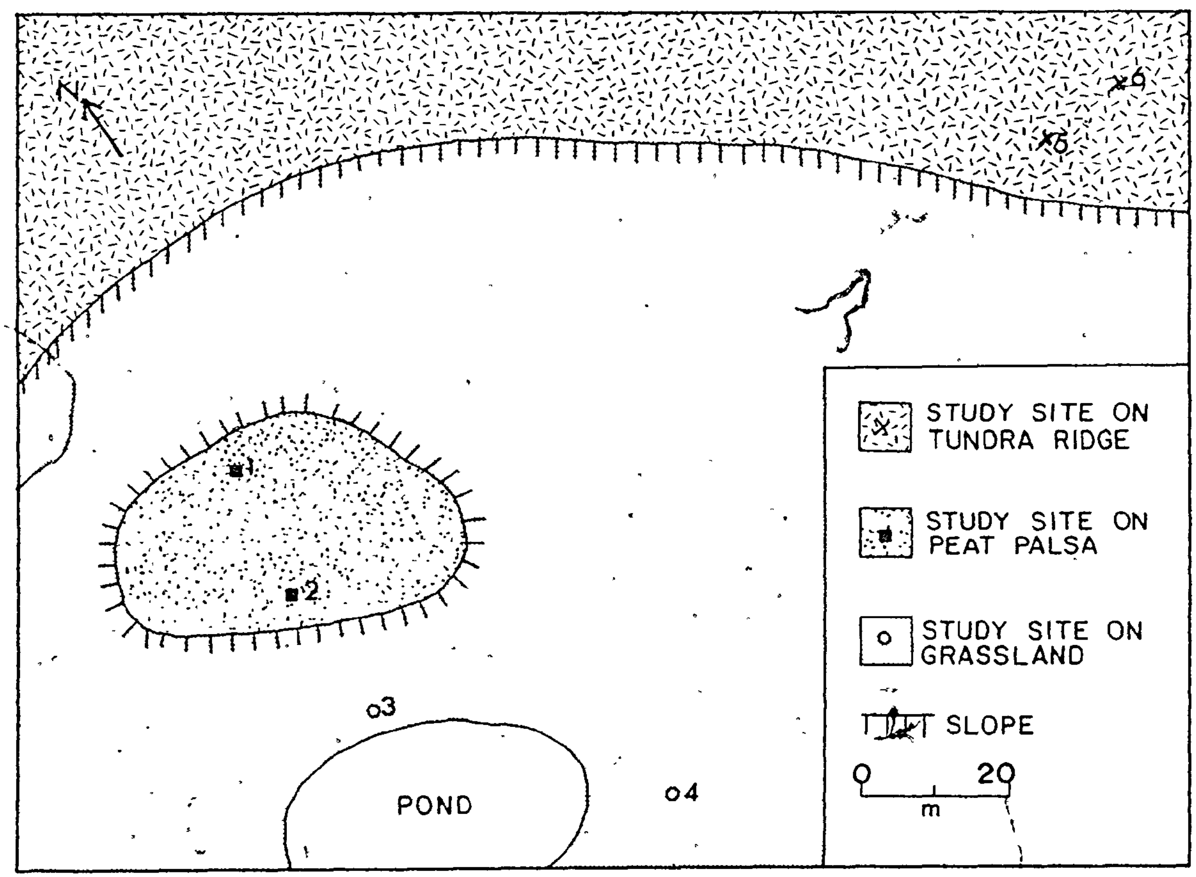
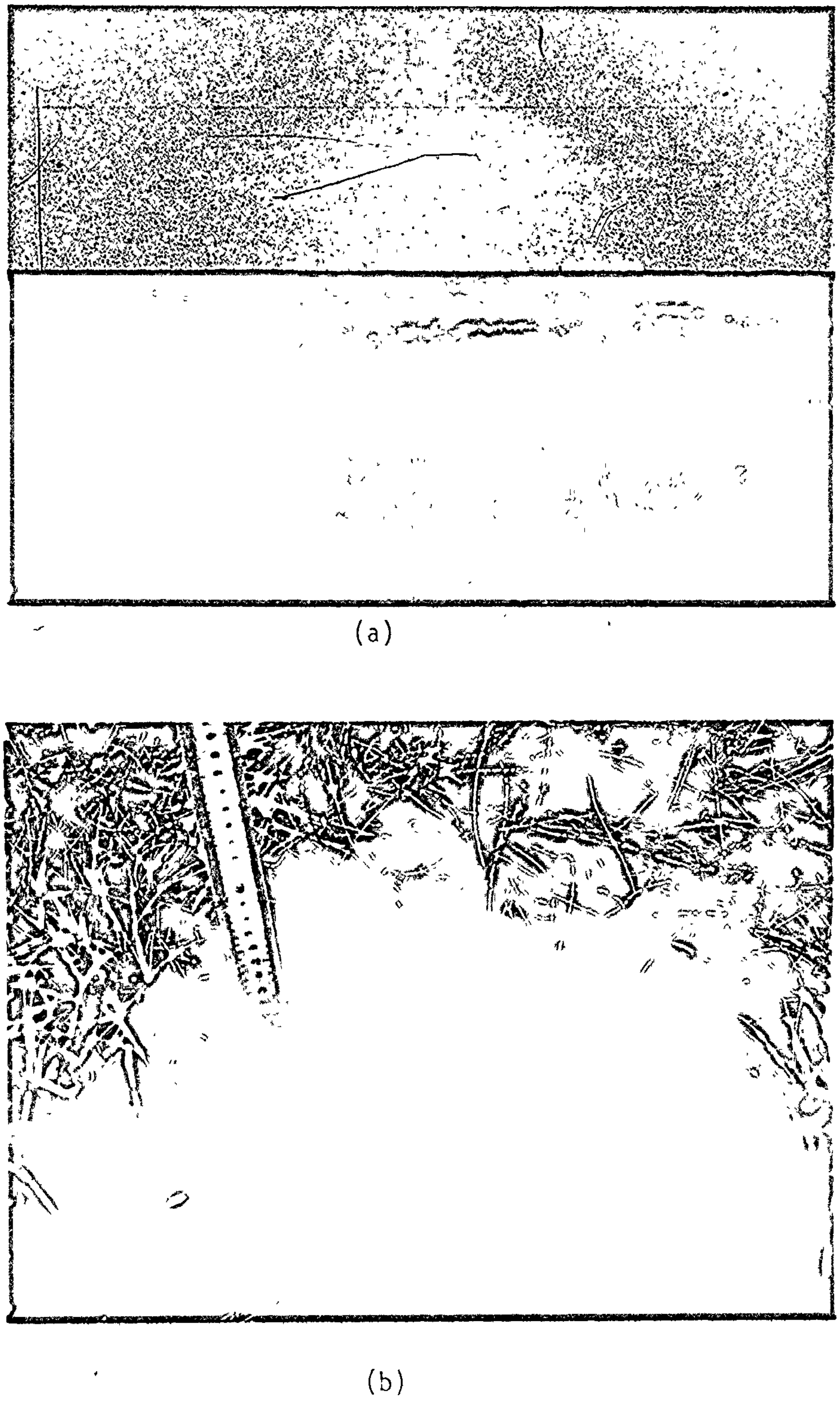


FIG 3.2 PLAN DIAGRAM OF STUDY SITE LOCATION



(a)

(b)

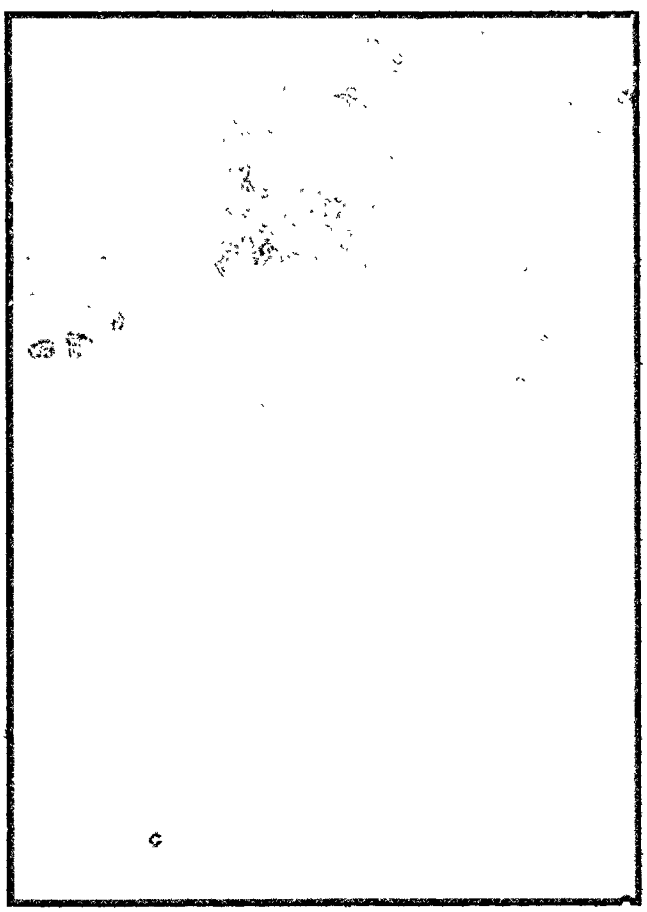
Fig. 3.3 Grassland Site

near saturation all season. Fig. 3.3 (b) shows a soil pit with a profile of 0.10 m thick layer of dense grassroots blending into 0.30 m of very wet, partly decomposed organics. This organic layer has a mean bulk density of 0.29 X 10^3 kg m^{-3}. At 0.40 m an abrupt change to a dense grey clay occurred. The water table was very near the surface as indicated by the rapid infilling of water in the excavated pit. Surface vegetation was dominated by sedges of Carex spp..

Protruding above the surrounding grassland is a small peat palsa. The relief varies from 1 m on the north edge to 0.5 m on the southwest. Surface cover consists of peat, lichens (Cladonia spp.), Labrador tea (Ledum groenlandicum), and numerous berry-producing plants such as bear-berry (Arctostaphylos rubra). The palsa is shown in Fig. 3.4 (a), with the soil profile illustrated in Fig. 3.4 (b). The solid, dense peat extended to the frost table. The extent of decomposition increased with depth but the whole profile consisted of a dark, black, fibrous mat with an average bulk density of 0.42 X 10^3 kg m^{-3}. This thick mat served as an effective insulator of the underlying permafrost.

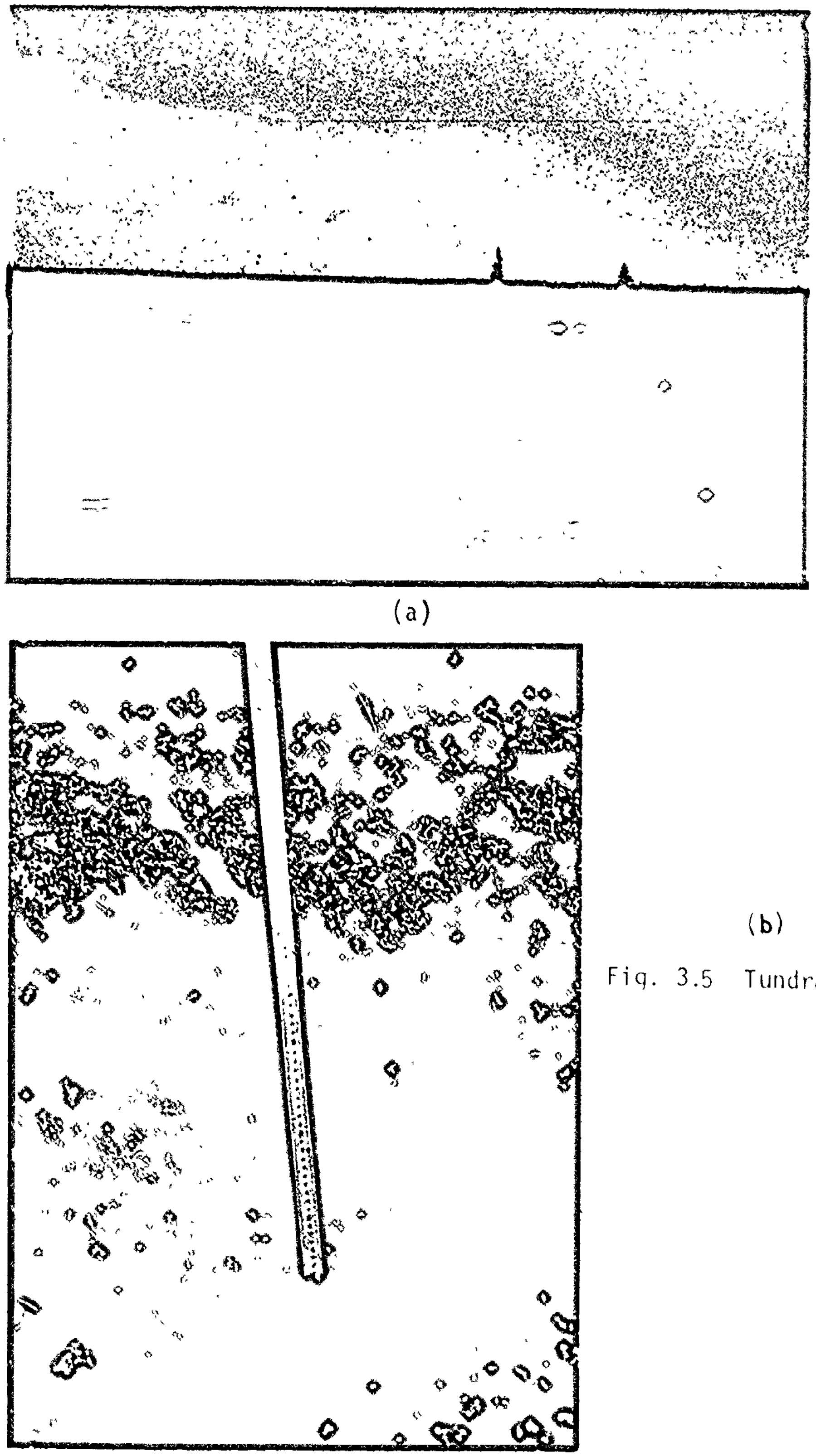
The remaining site characterizes a lichen heath habitat. Illustrated in Fig. 3.5 (a), it lies leeward of a small wooded area of black spruce (Picea mariana) and thus is protected from the prevailing northerly winds. Consequently, snow accumulation is deeper and remains longer than on the open, windswept tundra. The vegetation covering the coarse, sandy soils is composed of lichens (mainly C. Stellaris), arctic avens (Dryas integrifolia) and some sedges (Carex spp.). These plants are well-adapted to dry, harsh arctic conditions. Fig. 3.5 (b) reveals an organic layer less than 0.20m thick. It

(a)



(b)

Fig. 3.4 Peat Site



(a)

(b)

Fig. 3.5 Tundra Site

rapidly changes to a coarse sand of a mean bulk density of $1.39 \times 10^3 kg\ m^{-3}$. Intermittent layers of limestone pebbles are encountered.

The sites are indicative, both in vegetation and structure, of the low arctic of the Hudson Bay lowlands. Detailed understanding of the thermal processes in these terrains is necessary in order to typify thermal regimes in more inaccessible similar areas for mapping and planning purposes.

3.3 Instrumentation and methods

The primary aim of the research project necessitated that the instrumentation be mobile enough that 6 study sites surrounding a main research site could be monitored.

Ground temperatures were measured with a probe constructed by embedding a YSI Precision thermistor in the sharpened head of a 1.5 m steel rod graduated in 0.10 m intervals. Five temperature probes were assembled and individually calibrated against platinum resistance thermometers. Resistance changes of the thermistor were measured with an accurate digital resistance meter. Using this probe, soil temperatures were recorded at depths of 0.05, 0.10, 0.15, 0.20 and 0.30 m near the surface and at subsequent 0.10 m intervals through the deeper active layer to a maximum of 1.40 m for the tundra. The thermistors were well-suited to the programme because the large resistance changes for small temperature differences reduced errors due to connections or long lead lengths. Their ease of mobility permitted a novel measurement programme of soil temperatures in different terrain types.

Net all-wave radiation was measured continuously with a Swissteco pyrradiometer at a main control site, 300 m north of the study sites. At four periods during the season net, incoming and reflected solar radiation values were determined for the 3 land surfaces. The solar fluxes were measured with Eppley pyranometers on clear, sunny days and with this information characteristic albedo values were determined.

A short supplementary field season in July, 1980 facilitated a more detailed comparison of the net radiation over each surface. Continuous values of net radiation at identical time periods at a test site and the main site were needed to compare radiation balances. Accordingly a portable net radiometer was used to measure Q* over the peat surface for 65 hours with integrated half-hourly values recorded on a data logger. For each recorded value, an equivalent value for the main research site was measured. An identical procedure took place for 86 and 153 hours for the grassland and tundra respectively.

With these data, Q* Study Site was related to Q* Main Site by regression. Using the 12 values of daily total net radiation, the regression equation (Fig. 3.5) was

$$Q^* \text{ Peat} = Q^* \text{ Grassland} = Q^* \text{ Tundra} = .550 + .921\ (Q^* \text{ Main Site}) \qquad (14)$$

$$r^2 = 0.93$$
$$\text{standard error} = 0.77 \text{ MJ m}^{-2}\text{d}^{-1}$$

With this equation, the value of daily total net radiation for each day of the field season was determined.

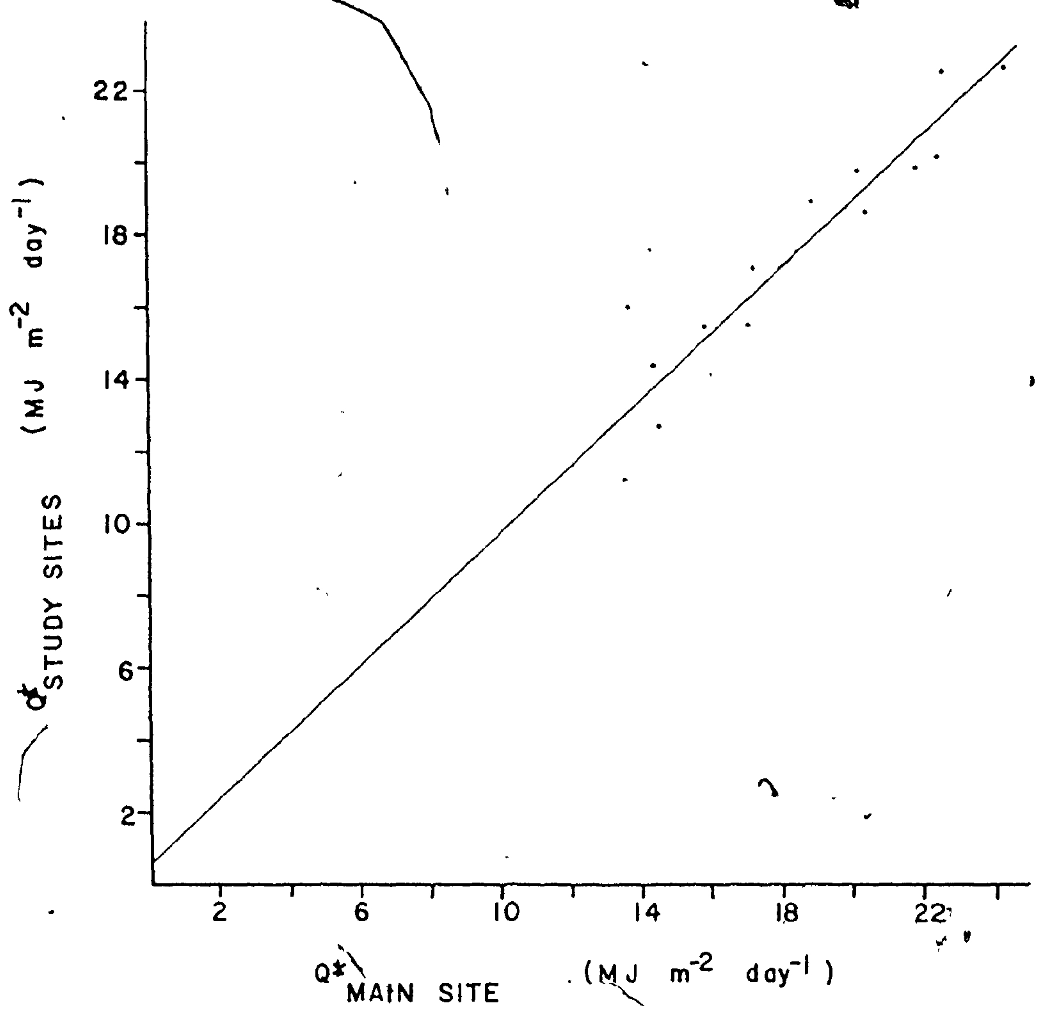


FIG 3.5 REGRESSION PLOT TO DETERMINE NET RADIATION OF STUDY SITES

Moisture content of the soils was determined as follows. Volumetric measurements were made four times during the field season on 28 June, 27 July, 10 August and 22 August. Sampling tins of 1.2 X $10^{-4} m^3$ were used to remove soil samples at 0.10 m intervals on a freshly excavated face of a soil pit. The net weight change of each oven-dried sample was divided by the volume of the tin to determine a volumetric measure. This provided soil moisture information for 4 discrete periods over the season. After graphing these data (Figs. 3.6) and assuming a constant rate of change between these dates, the intervening soil moisture values were interpolated.

An experiment performed in the summer of 1980 ascertained the saturation value and bulk density for each soil. These results concurred with similar field capacity determinations of the previous summer. The grassland saturates at a water content of 83% by volume. In the predominantly sand material of the tundra, the saturation value is 43%. This site is drier than both the grassland and the nearby peat site which becomes saturated at 63% water content.

The soil measurements extend only to the 0.40 m depth. In many cases, standing water or permafrost was encountered at or before this level. Late in the season, when the active layer was deeper than this, a constant soil moisture was extrapolated based on the homogeneity of the soil. At the actively thawing layer, the soils remain saturated, an assumption substantiated by the very wet nature of the soils.

Rainfall was measured daily at an adjacent open tundra site,

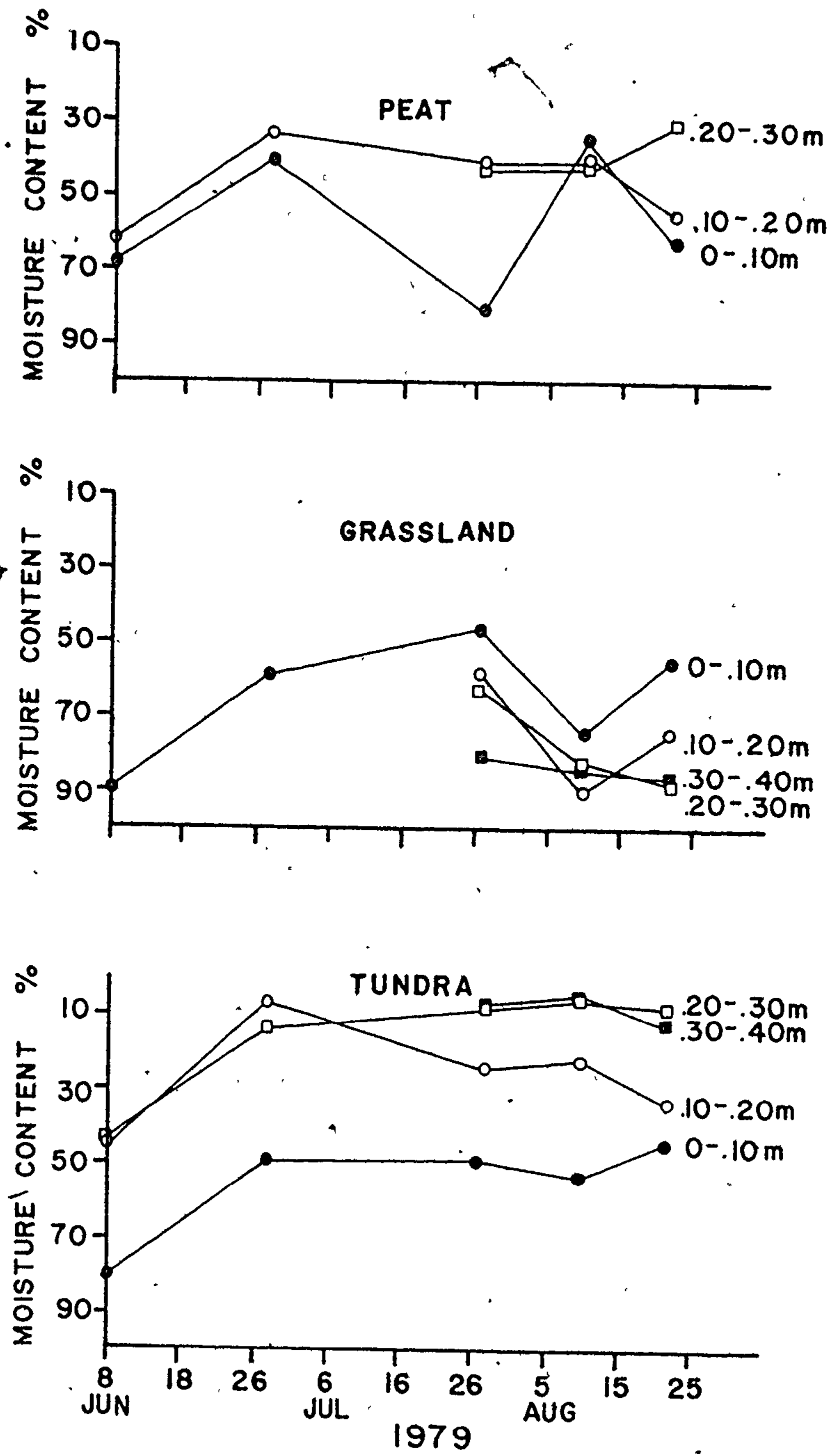


FIG 3.6 VOLUMETRIC MOISTURE CONTENT DURING THE THAW SEASON

using standard rain gauges sunk into the ground flush with the tundra surface.

Associated with each instrument and technique is a limited amount of experimental error which must be evaluated prior to discussion of the research results.

3.4 Measurement accuracy

For calculations of the soil heat storage term, soil temperatures and heat capacity were required. The precision thermistors used to measure soil temperatures were accurate to ± 0.1 C. Since measured temperatures ranged mainly between 1C and 20 C, percentage errors were between 0.5% and 10 %. The resistances were measured with a Kiethley multimeter which was accurate to ±1%. Hence maximum measurement errors for temperatures range between 1.5% and 11% depending on the temperature. Here it is assumed that a maximum error of 5% is representative.

A second source of error is associated with the determination of heat capacity. Here the accuracy of the soil moisture content is important. An estimate of error in the field sampling and in laboratory analysis is ± 10% of the volumetric moisture. Analyzing the sensitivity of C to moisture shows that for a 10% increase in moisture, heat capacity increases up to 5%. This is considered an appropriate error for the heat capacity term. Thus the total maximum error in the determination of the sensible heat storage is 10%. The root mean squared error estimate is 5%.

This soil heat component is included in the calculations of the evaporation term of the energy budget. Following eq. (11), total error in Qe will reflect errors in α, Q* and Qg, such that

$$\frac{dQe}{Qe} = \frac{d\alpha}{\alpha} + \frac{dQ^*}{Q^*-Qg} - \frac{dQg}{Q^*-Qg} \qquad (15)$$

α is accurate to $\pm$ 10% based on studies by Rouse (1981 b) and Stewart and Rouse (1976). Net radiation measured with a Swissteco pyrradiometer is correct to $\pm$ 10% also. Further error of 10% results by obtaining Q* by regression. Since Qg/Q* has an average value of 0.15, $dQ^*/(Q^*-Qg) \simeq 0.02$. With these values, the maximum error in Qe is close to 32%. The root mean squared error of Qe reduces to $\pm$ 13%.

The water balance calculations of soil moisture and subsurface horizontal transfer depend on the accuracy of the input variables of precipitation and evaporation. Rainfall is reliable to $\pm$ 5% and the accuracy of the evaporation term has been evaluated at $\pm$ 13% r m s e. The error in the net change in soil moisture content has been established as $\pm$ 10%. Using these values, error may be determined with the expression

$$d\Delta Wz = 0.1\Delta Sm - 0.5\Delta P + .13\,\Delta E \qquad (16)$$

The maximum error associated with the tundra ranges from 30 to 80 mm with an average relative error of 46%. Similar values for the peat and grassland are 20 to 80 mm with a mean relative error of 29%, and 30 to 60 mm with a mean relative error of 24% respectively in the subsurface water movement.

CHAPTER IV

Results

4.1 Net radiation

Fig. 4.1 shows the variation of daily net radiation. Short term fluctuations are due to cloud or fog conditions. The lowest radiation receipt (1 August) occurred on a day with 20 mm of rain. The maximum net radiation occurred on 23 June, two days past the summer solstice.

Calculated albedo values should be similar as all surfaces were naturally vegetated. The grassland albedo of 0.13 is relatively low due to patches of standing water and the very wet soils. The drier peat has an albedo of 0.15 and is more characteristic of a vegetated surface. The tundra is covered in light coloured lichens and sand. Accordingly, the albedo is slightly higher at 0.16. These sites are initially differentiated on the basis of their albedos.

4.2 Soil moisture and heat capacity

The soils remained wet throughout the season. The seasonal pattern of soil moisture changes is shown in Fig. 4.6 where the volumetric moisture content has been converted to a water depth equivalent.

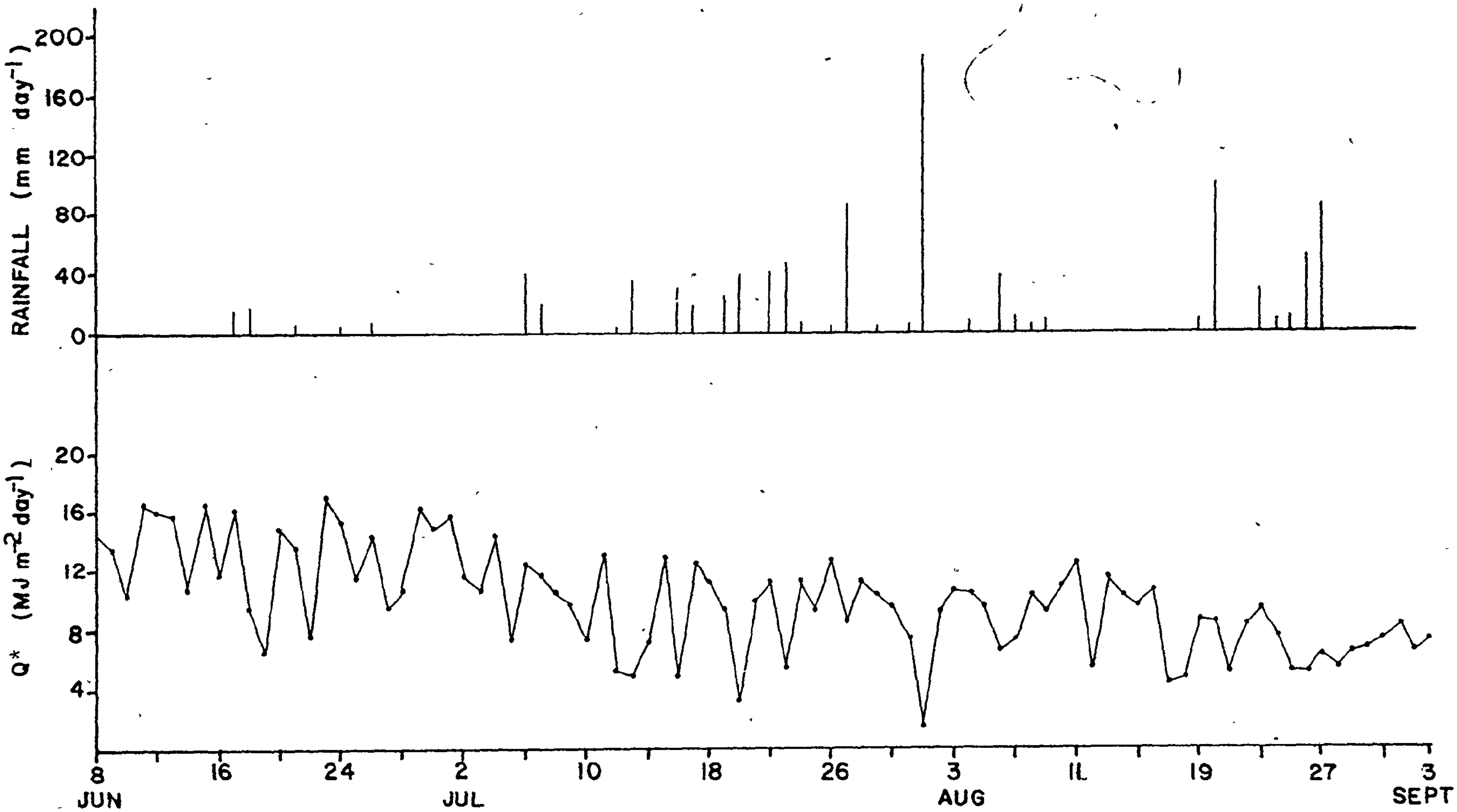


FIG 4.1 TOTAL DAILY NET RADIATION AND RAINFALL

Following snowmelt the soils gradually dried until August when the soils became wet following rainstorms on 31 July, 1 and 4 August. The grassland differs most noticeably from the tundra in its soil moisture characteristics.

The tundra site situated in the lee of a small woodlot has a deep and long-lasting snowcover. As a result, soil moisture is higher than the surrounding more exposed tundra. Unlike the grasslands, the tundra surface soils remain wet due to the coverings of lichens and organics. In the upper 0.20 m volumetric moisture content approached 50%. At 0.40 m soil moisture decreased to 10%. In contrast, the grassland which experienced some surface drying remained saturated below a depth of 0.20 m. The moisture content of the thawed peat ranged between 80% and 32% in the 0-0.10 m. layer and 30%-40% in the deeper layers.

These soil moisture values were used to calculate heat capacity for each 0.10 m layer in the thaw zone of each site (Appendix 1). Below 0.40 m, the soil was assumed to be homogeneous and the heat capacity was set at a constant value equivalent to the value at 0.30 m. The average heat capacity was largest for grassland (3.82 $MJm^{-3}K^{-1}$), due to high organic and moisture contents. This is followed closely by the peat with a mean heat capacity of 3.37 $MJm^{-3}K^{-1}$. The heat capacity of the tundra (2.67 $MJm^{-3}K^{-1}$) is smaller because of less organics in the soil and smaller moisture content.

4.3 Energy Balance

The energy balance of the three sites for the period 8 June to 3 September is summarized in Table III. The net radiation was identical

	GRASSLAND	PEAT	TUNDRA
ΣQ^* (MJ m^{-2})	864	864	864
$\Sigma \Delta S$ (MJ m^{-2})	159.1	120.0	109.9
$\Sigma \Delta S / \Sigma Q^*$	.18	.14	.13
ΣQ_e (MJ m^{-2})	509.8	403.5	408.3
ΣQ_h (MJ m^{-2})	195.4	340.8	346.1
L_f (MJ m^{-2})	148.0	113.2	106.8
$\Sigma L_f / \Sigma \Delta S$	.93	.94	.97
α	.13	.15	.16
K (x $10^6 m^2 s^{-1}$)	.21	.12	.48
z (m)	.56	.48	1.65

TABLE III: Seasonal Energy Budget

for each because of the use of a common regression equation (eq. 14) to calculate the flux from measurements at the main site.

In this permafrost region, the magnitude of the soil heat storage term, S, is very important since the thawing process depends on this energy. The change in soil heat, ΔS, was determined for each depth increment of 0.10 m. The energy stored in the full active layer was obtained by totalling these incremental values. The daily variation of ΔS during the thaw season (Fig. 4.2) was greatest for tundra. It also experienced three periods when heat from the soil was lost to the atmosphere. This resulted from steep near-surface temperature gradients, high conductivity and low buffering ability of the soil due to the low heat capacity. The peat and grassland absorbed soil heat at a more constant rate. The greater moisture content of the grassland resulted in greater heat storage. It is appropriate that the largest heat storage (10 MJm^{-2}) for a measurement period of 3 days occurred in the grassland. Soil heat storage in the grassland comprised 18% of the net radiation. Storage values of the peat and tundra are similar and the ratio of $\Sigma\Delta S/\Sigma Q^*$ was 14% and 13% respectively. Although the tundra received an equal energy input, proportionately less was consumed as soil heat. Rouse (1981a) finds a similar soil heat storage (18%) in a raised beach tundra site. These studies suggest that the soil heat storage component of the energy balance is important in permafrost.

The soil heat quantity has a latent and sensible heat component. The latent heat is dominant (Table III) and contributes about 95% of the total heat storage. This leaves only about 5% as sensible heat to

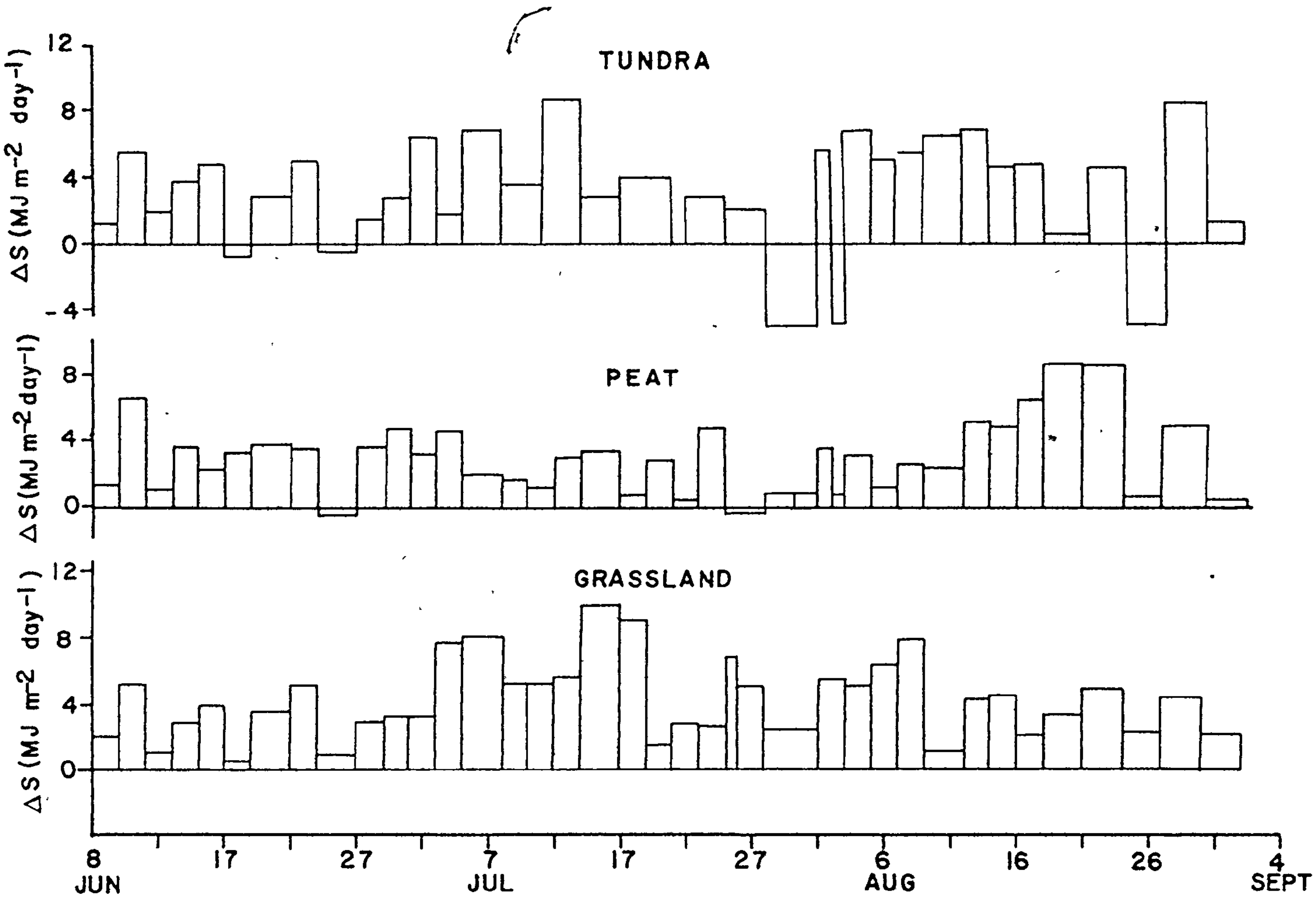


FIG 4.2 DAILY CHANGE OF SOIL HEAT IN THE ACTIVE LAYER

raise the soil temperatures. At the start of the measurement programme the thaw depth was 0.15 m in each site. By the end of the study, thaw depths differed substantially between terrain types. Active layer depths (Fig. 4.3) reached a maximum of 0.56 m and 0.48 m in the grassland and peat and at least 1.65 m in the tundra. As noted earlier, the thermal properties in the ground controlling the variable active layer depths are thermal conductivity and diffusivity.

Thermal conductivity is difficult to measure as it varies with composition, temperature and moisture content. The thermal conductivity values presented, and used, are averages for these subsurface materials. The high silica content and relatively low moisture content of the tundra produces the highest conductivity (1.28 $Wm^{-1}K^{-1}$). The saturated grassland has a conductivity of 0.80 $Wm^{-1}K^{-1}$. The combined influences of high moisture content and high porousity reduces the conductivity of peat to 0.42 $Wm^{-1}K^{-1}$. This agrees well with the value of 0.53 $Wm^{-1}K^{-1}$ for the thermal conductivity of peat found in the Solenyi area of the Soviet Union (Pavlov, 1981).

This low conductivity explains in part the effective insulating ability of the peat which results in a shallow active layer. The conductivity of both air and water is low (Table I). Consequently during the summer, the permafrost of a peat hummock is protected from excessive heat gains. The reverse occurs in the winter when ice, with its high conductivity supplants the water and air. Large losses of heat from the peat enhances the frozen conditions of the subsurface material.

The second soil thermal characteristic is the thermal diffusivity, K, the ratio of thermal conductivity to heat capacity, $K = \lambda/C$. It measures the time for temperature changes to travel through the soil.

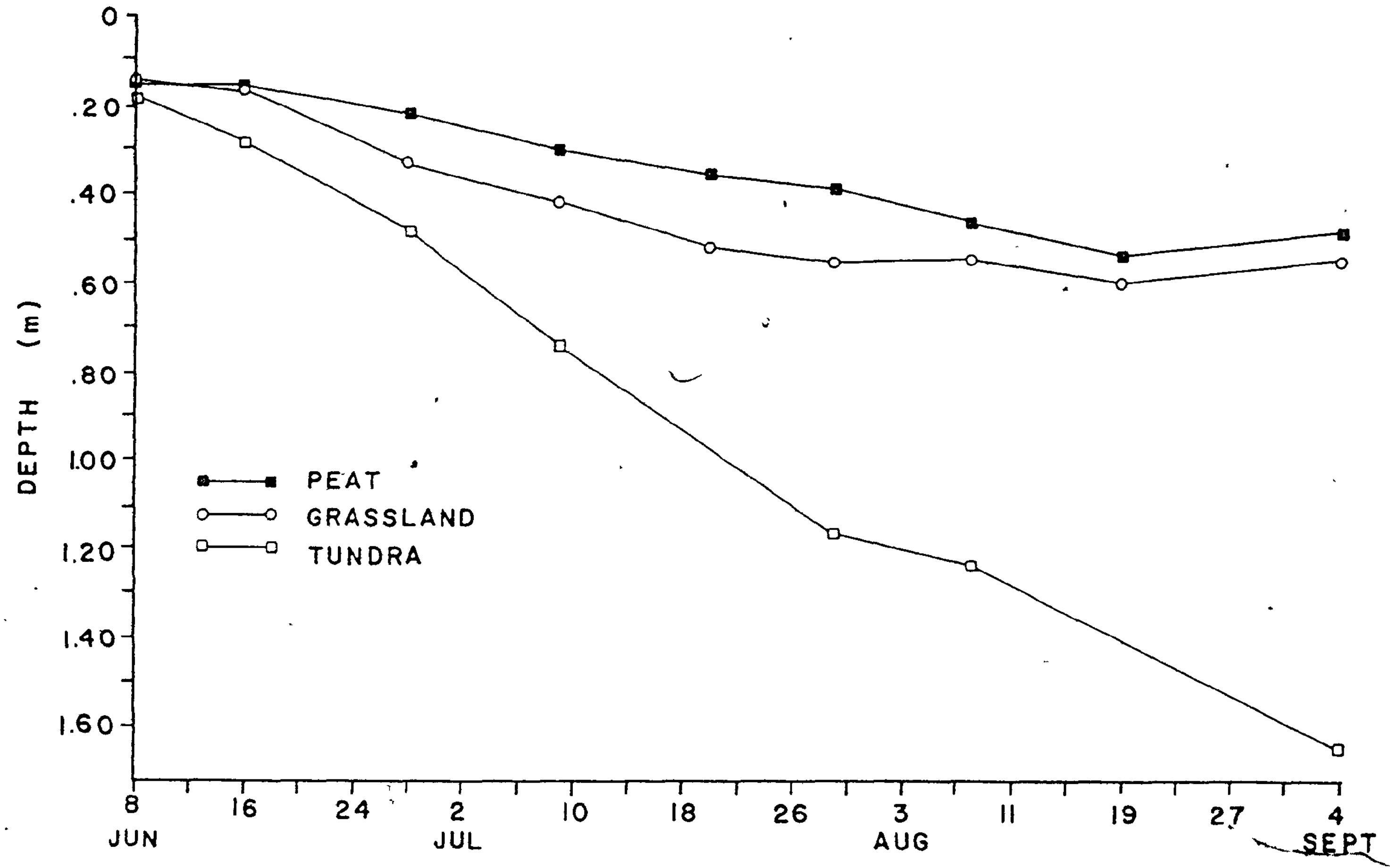


FIG 4.3 ACTIVE LAYER DEVELOPMENT

The peat has the lowest diffusivity of 0.12 X $10^6 m^2 s^{-1}$ compared to the 0.21 X $10^6 m^2 s^{-1}$ for the grassland. The tundra diffusivity of 0.48 X $10^6 m^2 s^{-1}$ is four times greater than the diffusivity of the peat. Thus the tundra has the deepest active layer since it diffuses heat energy more efficiently than the peat or grassland.

The temporal pattern of the total increase in soil heat storage (Fig.4.4) differs between terrain types. Initially, all surfaces absorb energy equally. Beginning on 4 July the stored soil heat of the grassland increases rapidly for a three week period, the rate diverging from the rate for peat and tundra. This period corresponds to the large heat energy gains between 2-18 July (Fig.4.2). For the remainder of the season a steady rate of accumulation is maintained. Tundra and peat were similar throughout the season. This is substantiated by the similarity of the $\Sigma\Delta S/\Sigma Q^*$ ratios. Both surfaces absorb less energy than the grassland. The rate of heat storage for tundra and peat increased in early August but failed to attain the grassland value. The increase in total heat storage for peat and tundra is 75% and 60% of the grassland value.

The remaining terms of the energy balance are the latent and sensible heat fluxes. The results are presented in Table III and diagrammatically in Fig.4.5. This figure shows the similarity of the energy balances of the peat and tundra terrain. Grassland is different since evaporation increases to 59% of Q* and sensible heat loss increased to almost one-half that of the other terrains.

These results differ from those of Weller and Holmgren (1974). During their short post-melting period, evaporation, sensible heat and soil heat comprised 73%, 18% and 9% of Q* respectively. The

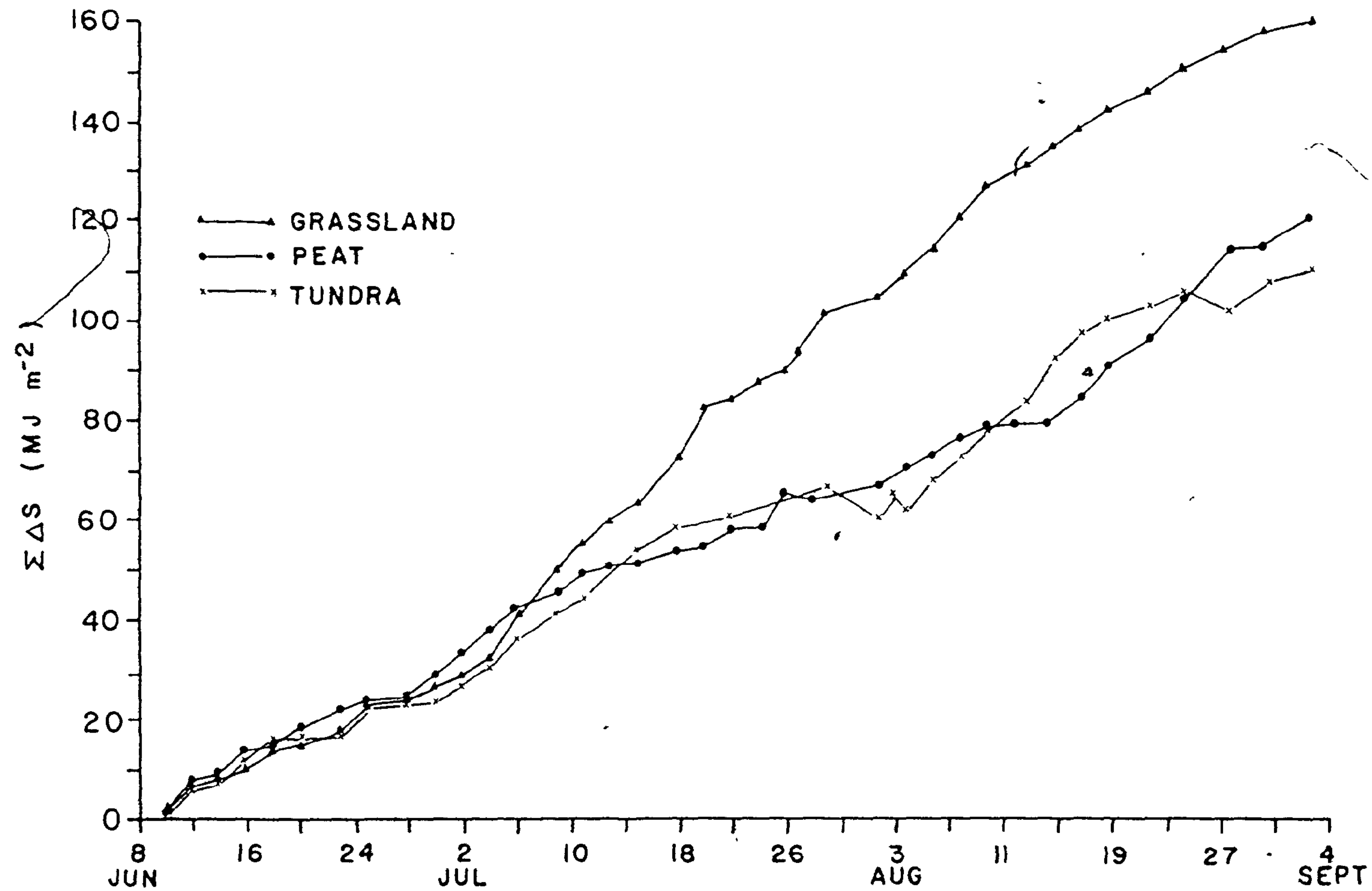


FIG 4.4 TOTAL INCREASE IN SOIL HEAT STORAGE

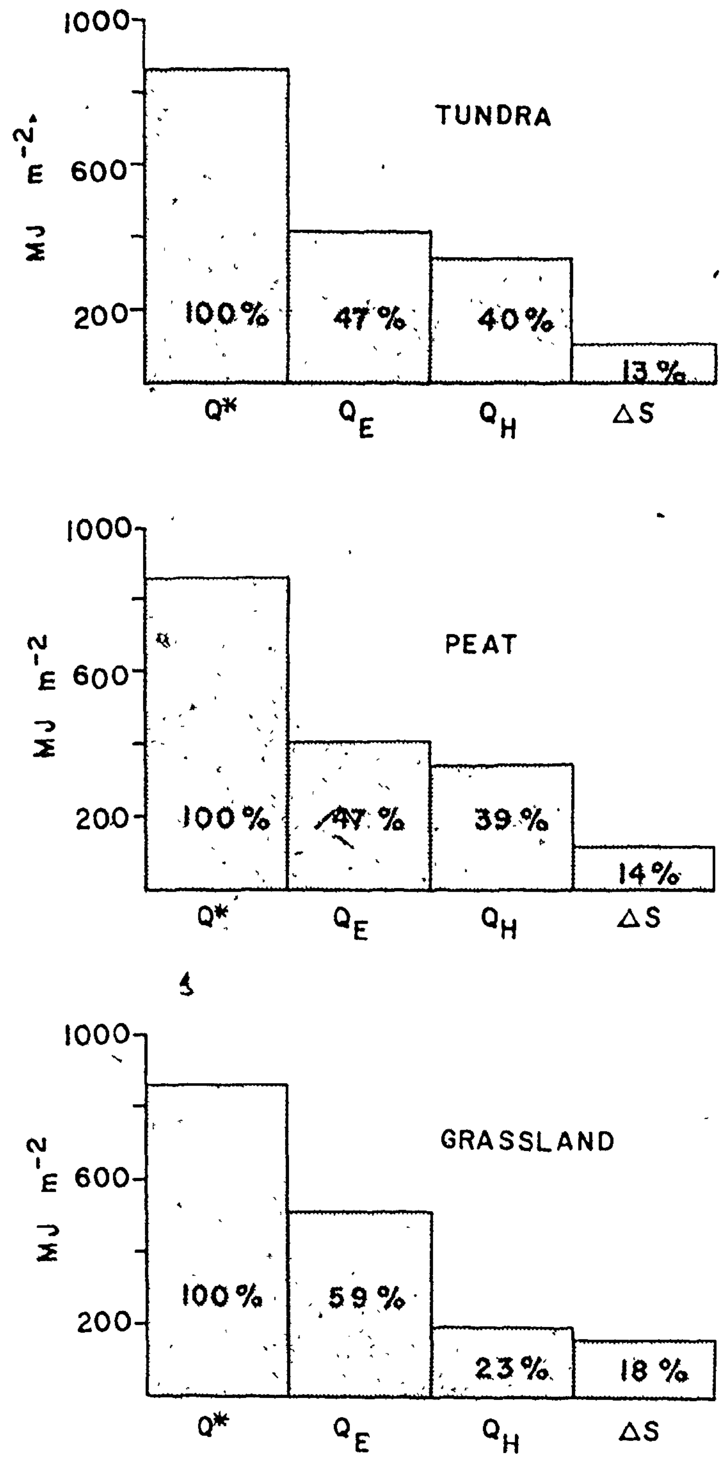


FIG 4.5 SEASONAL ENERGY BUDGET
8 JUN - 3 SEPT

energy was repartitioned during the two-week midsummer season where the respective fluxes were 66%, 32% and 2% of Q*. The most noteworthy discrepancy is the magnitude of the soil heat flux. In their study, no mention of moisture content is offered and the heat transfer through convection of water is ignored making the comparison of results difficult. The soils of Barrow, Alaska contain substantially less organic material and are generally drier than those around Churchill. Because of these differences, the soil heat term will be larger in the study sites' soil and this term will include a large latent heat component.

In permafrost areas, the soil heat energy thaws the subsurface producing an active layer of variable depth. The associated release of soil water modifies the thermal regime and further helps to alter the regional energy and water balances.

4.4 Water Balance.

Seasonally total values (summations of weekly values) of the water balance for each terrain type appear in Table IV.

Precipitation (~100 mm) was evenly distributed throughout the summer with measurable rainfall occurring on 38% of the field days.

Evaporation reached maximum values in June when sites were wet and greatest amounts of energy were available. Later in the season evaporation decreased as values of the net radiation declined.

Soil moisture changes in the active layer, ΔSm were virtually constant throughout the thaw season for grassland, whereas the tundra

	Grassland	Peat	Tundra
Precipitation (mm)	99	99	99
ΔSm (mm)	-10	-70	-311
E (mm)	200	158	160
ΔWz (mm)	91	-11	-250

Table IV Seasonal Water Budget
8 June - 3 September

sustained a large depletion of 311 mm. The peat experienced a smaller loss of 70 mm. The pattern of thaw varies between sites (Fig. 4.6). Three periods can be distinguished. Before July, thaw was gradual and total soil moisture decreased. For most of July the rate of thaw accelerated. For the tundra and grassland this was a period of rapid soil water loss. For the tundra, 50 mm of water was lost in each of 4 consecutive weeks in July. Soil moisture for peat increased slightly at this time. In the final period (after July) the rate of thaw decreased as the active layer reached its maximum. Little further moisture loss occurred in the peat or tundra while soil moisture increased at the grassland site.

Variable soil moisture changes are largely a response to the subsurface water flux term, ΔW_Z. This term was calculated from eq. (14) for weekly and seasonal intervals (Table IV). The low-lying grassland gained 91 mm through subsurface flow while the peat and tundra sites lost water through this subsurface flow. In particular, the tundra lost water rapidly at an average rate of 3.0 mm day $^{-1}$, compared to 0.2 mm day^{-1} for the peat and -11.0 mm day $^{-1}$ for the grassland.

The implication of this subsurface water flux is examined in the next chapter and related to the energy and water balances of the three terrain types under investigation.

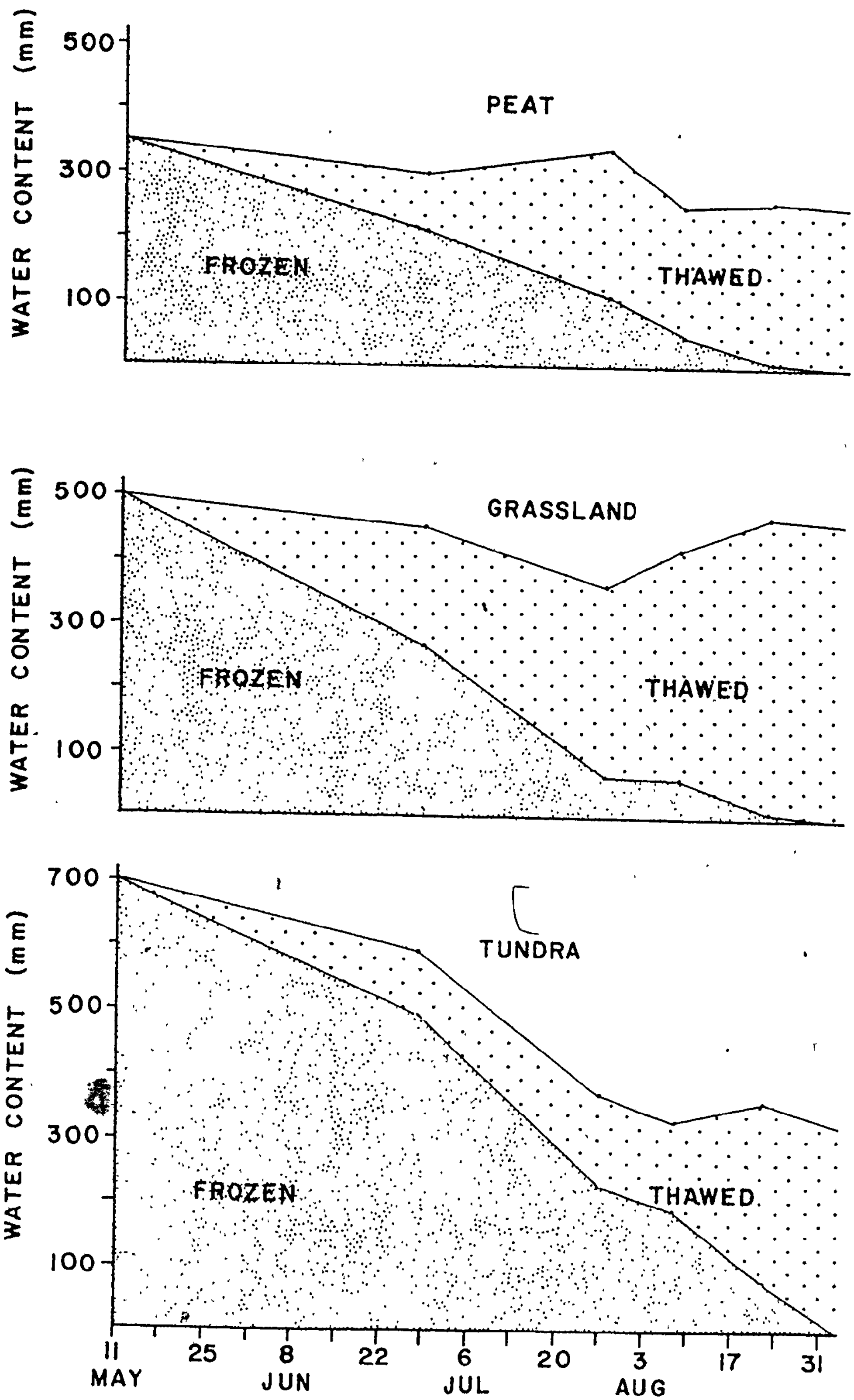


FIG 4.6 TOTAL WATER CONTENT DURING THE THAW SEASON

CHAPTER V
Discussion

This description of the soil heat and water balances of 3 examples of northern terrain contributes to the very limited data base concerning permafrost genesis, distribution and thermal regime. Few studies have concentrated in the Hudson Bay Lowlands or have as lengthy a study period. In addition, the water balance of permafrost regions is scantily documented; however, the annotated bibliography by Dingman (1973) summarizes the research published prior to 1970. Several important observations arise from this analysis.

5.1 Soil heat storage

Past energy budget studies in the northern latitudes have emphasized the magnitude of the surface energy fluxes. As a result there are few studies with which to compare the soil heat storage term of this project. The surface soil heat flux measures only the amount of energy crossing the surface interface. Since the heat flux across the frost table is unknown, the storage term becomes an estimate of the soil heat flux for a given period. This condition allows the present results to be compared to previous arctic and subarctic research.

The soil heat storage for each terrain type is much larger than reported in other arctic studies. For example, Smith (1975b) reporting on a limited set of measurements from the Mackenzie Delta found the ground heat flux to be 8% of the net radiation. His calculations of heat storage were based on basic heat conduction theory and no mention was made of the volumetric moisture content of the soils. In his study and others, a basic assumption is that conductivity remains constant through the measurement period. This assumption may account in part for the discrepancies between the present study's value of soil heat of 13% to 18% compared to the reported 8%.

Thompson and Fahey (1977) working on Broughton Island, found the soil heat flux, Q_g, averaged 7% of Q* during a midsummer period. In 6 of 9 sites, Q_g ranged from 5 to 11% of Q*. These values indicate that the energy available for soil heat is limited. Much more energy is expended in evaporation and sensible heat. Again, a constant thermal conductivity was used. The one-dimensional conductive model used, ignores mass transport due to water vapour and liquid flow. Thompson and Fahey comment that Bowen ratios and energy balance calculations do not fully explain differences in thaw lines, stressing the need for more information on the ground-soil thermal regime. The deepest active layer of 0.8 m was found in moist sandy soils. This agrees with the results of the present study.

Addison and Bliss' (1980) results for King Christian Island, a polar semi-desert, found a wide range of Q_g values varying from 4% of Q* to 14%. At this site, Q_g correlated with the depth of the insulative organic layer such that the greatest Q_g values were associated with lichens and bare rocks.

Two additional studies of soil heat and water balances in the

Hudson Bay lowlands are by Rouse (1981 a, b) which include data extending through two consecutive thaw seasons. The soil heat component of the energy balance was calculated in a manner similar to this study. The magnitude of soil heat storage was 18% of Q*, similar to the values of 18%, 14%, and 13% of Q* in the grassland, peat and tundra sites. These sites remained wet throughout the season. Hence, soil heat is a major componet of the energy balance in wet terrain.

5.2 Subsurface water movement

The ΔW_z term (Table IV) for both the tundra and grassland points to a substantial subsurface flux of soil water as shown in Fig. 5.1. The tundra experiences a substantial water loss through seepage which would flow along the topographic and frost table gradient to the lowlands. The tundra soils are initially near saturation due to a deep, long-lasting snowcover. As melt is initiated, water ponds on the frost table then begins to seep along the frost table as "suprapermafrost groundwater" (Wright, 1980). The period from Week 2 to Week 8 witnesses a rapid seepage loss averaging 7.0 mm day^{-1}. This coincides with a period of rapid active layer development.

The grassland pattern shows little net subsurface flux until late July after which the flux increases quickly. This reaches a maximum by Week 11, corresponding to the declining water flux in the tundra. This evidence suggests a subsurface flow from the upland tundra to the low-lying grassland with a significant time lag experienced.

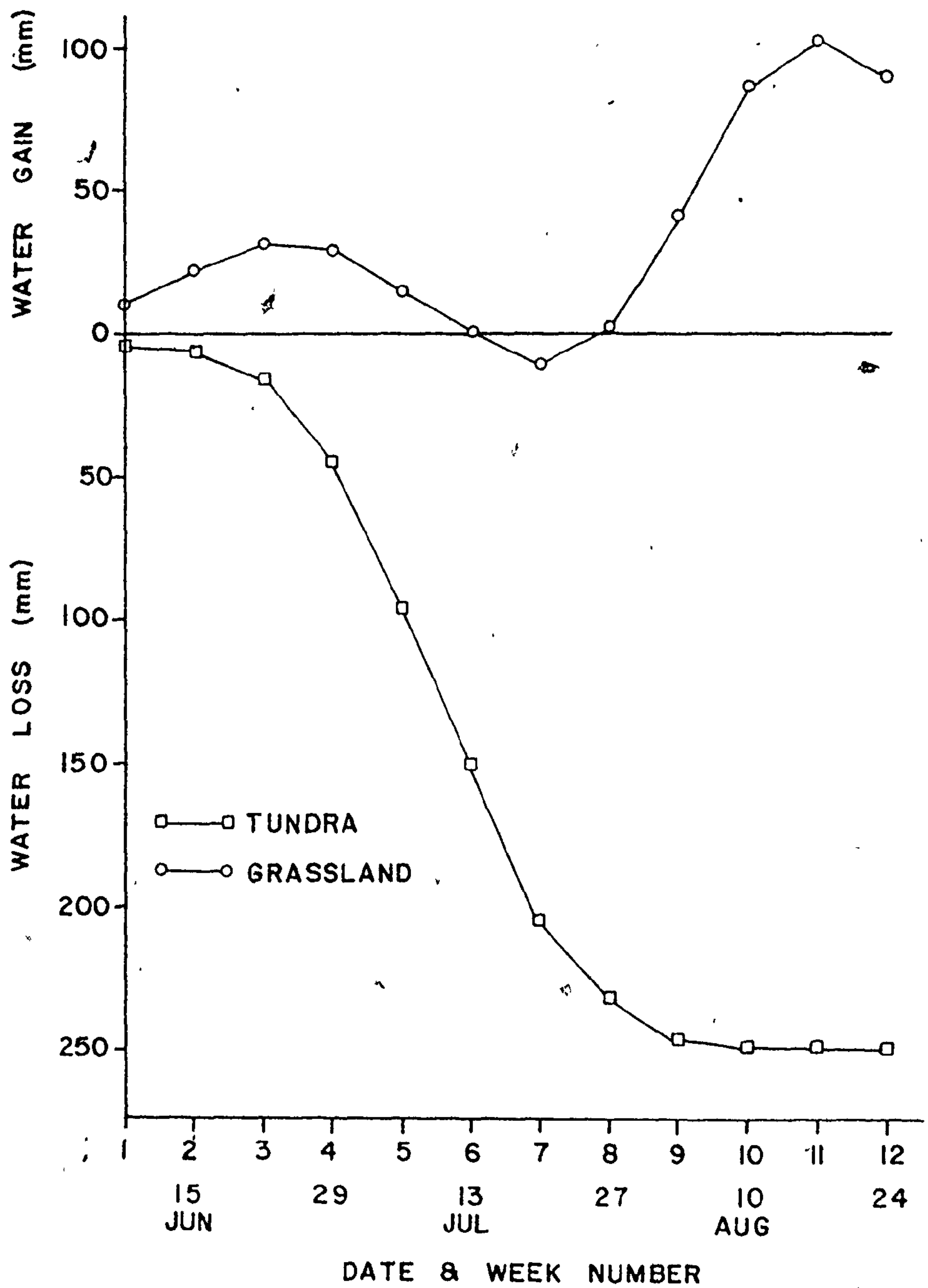


FIG 5.1 CUMULATIVE SOIL WATER LOSS

The concept of subsurface water movement in both frozen and thawed soils has been considered by Wright (1980) for areas near Schefferville, Quebec. This research on the water balance of a subarctic lichen heath ridge confirms that the theories of purely conductive heat transfer are inadequate to explain the observed temperature patterns and thaw depths. Wright proposes that significant quantities of heat are transported downward via water seepage from the thawed material into the still-frozen soil only to be released at a future time as latent heat when the thaw line reached this depth. As pointed out by Wright, the importance of latent heat transfer extends to the lateral concentration of heat by suprapermafrost groundwater flow. Wright's work supports the present hypothesis of lateral subsurface water movement, affecting active layer depths.

The need for quantitative data on the thermal and moisture regimes in arctic soils was recognized by Luthin and Guymon (1974) who examined the soil moisture - vegetation - temperature relationship in Alaska. To explain the observed temperature patterns, they proposed an "unmeasured lateral seepage from higher lands" (Luthin and Guymon, 1974). The suggested scheme places topographic drainage in the paramount position of control in the model. Poor drainage will induce a vegetation cover including peat which affects the thermal regime of the mineral soil, and permafrost results if a freezing temperature is maintained.

A strong linkage in the hydrologic characteristics of the sites is indicated by Fig. 5.1, and is due to the large subsurface water flux. The depth of active layer development is not only a response to the normal heat flux but also to the thermal imports and exports associated with the subsurface water fluxes. Considered in this manner, this water flux acts as an advective source of heat energy, supplementing the conductive soil heat component. In turn, these water fluxes are strongly affected by varying active layer depths in the terrain units throughout the thaw season.

CHAPTER VI

Conclusions

6.1 Summary and Conclusions

Near-surface thermal regimes in different permafrost terrain types were studied at a research site near Churchill, Manitoba. The study of the pattern of thaw in different terrain types as it relates to the solar energy received and the mechanisms of ground heat transfer encouraged further investigations into the interaction of the heat and water balance on short term effects of water movement in active layer development. In zones of continuous permafrost, the extent of the active layer controls the overall hydrological and biological patterns.

These patterns become complex in the varied landscape of the Churchill region. Three distinctly different terrain units allowed a comparative study, showing the importance of terrain type in regulating the surface energy balance. The tundra terrain, with a surface vegetation of lichens, grasses, peat and stunted black spruce covers a wet sandy soil beneath a 0.20m organic layer at the surface. This differs from the remaining two terrain types which are composed totally of organic material. Occupying the topographically low areas, the grassland supports a vegetation covering of sedges. Approximately 0.50 m of decomposing organic material of peat and grass roots overlies a layer of marine clays.

A peat palsa, a common landform in the north, supports some small plants and shrubs and several lichen varieties. The insulating abilities of peat prevent a deep active layer from developing and in this case limits it to less than 0.5m.

The methodology for the project was simple to facilitate mobile temperature measurements. For 3 months of thaw, soil temperatures and active layer depths were monitored every second day. To complement these data, net radiation also was recorded and found to be the same for each terrain type.

With the calculations of a soil heat storage term, the remaining energy balance components of latent and sensible heat were derived. Results demonstrate that the dominant midsummer energy flux is evaporation which comprises 59% of Q* in the grassland and 47% of Q* in both the peat and tundra sites. The sensible heat flux made up 40%, 39% and 23% of Q* in the respective three sites. The soil heat term constituted 18% of the available energy in the grassland and 14% and 13% of Q* in the peat and tundra, thus confirming the important role of soil heat in permafrost.

Cumulative soil heat values show that all terrains do not absorb heat energy equally; the soil heat storage of the peat and tundra equalled only three-quarters of the grassland value. Subdivision of the soil heat term showed that the major component is the latent heat required to melt ice in the permafrost. This comprises 93% to 97% of ΔS in the peat, grassland and tundra. The water released by this melt process contributes to the very wet soils and produces a subsurface water flux.

The subsurface water movement was estimated from rainfall, evaporation and a change in soil moisture context. The lateral drainage of water along the frost table from the tundra to the grassland suggests a hydrologic interdependence between adjacent terrain types. It also dramatizes the importance to the regional water balance of varying thermal and soil moisture regimes in permafrost terrain.

6.2 Future Research

The important goal of future research pertinent to the water balance of permafrost areas is the evaluation of the annual pattern and rates of soil moisture movement. Preliminary research of the soil moisture-temperature regimes in Alaska indicate that moisture movements and redistribution are most dynamic in the winter (Guymon, 1976; 1975). Total year round data would contribute to the sparse information on the spatial variability of the ice content of the permafrost and the water content of the active layer.

There is need to focus on the problem of quantifying the thermal processes of the "buffer zone"; that zone between the thermal dynamics of the atmosphere and the inorganic subsurface materials. These simulations depend on parameterizing the surface energy budget which in itself is subject to the vagaries of the climate.

Ultimately the aim is the ability to predict the extent of shallow permafrost and the attendent soil heat and water balances based on easily measured quantities including net radiation, temperatures and soil characteristics. The impact of land cover modification, either intentional or accidential, on the surface energy balance may be evaluated and provide valuable insights for planning purposes.

REFERENCES

Abbey, F. L., D. M. Gray, D. H. Male and D. E. L. Erickson, 1978. Index models for predicting ground heat flux to permafrost during thaw conditions. Proc. Third International Conf. on Permafrost. Vol. 1. Edmonton, Alta. pp.4-9.

Addison, P.A. and L.C.Bliss, 1980. Summer climate, microclimate, and energy budget of a polar semidesert on King Christian Island, N.W.T., Canada. Arctic and Alpine Res. 12(2): 161-170.

Brown, R. J. E., 1978. Influence of climate and terrain on ground temperatures in the continuous permafrost zone of northern Manitoba and Kewatin District. Proc. Third International Conf. on Permafrost, Vol. 1, Edmonton, Alta., pp. 15-21.

__________ and T. L. Péwé, 1973. Distribution of permafrost in North America and its relationship to the environment: a review 1963-1973. North American Contribution to Second International Conf. on Permafrost, Yakutsk, Siberia 1973, NRC publ. 13874.

Brown Beckel, D. K., 1957. Studies on the seasonal changes in the temperature gradient of the active layer of soil at Fort Churchill, Manitoba. Arctic 10 (3): 151-183.

de Vries, D. A., 1963. Thermal properties of soils. Physics of a Plant Environment. Van Wijk (ed.) Amsterdam: North Holland Publ. Co. pp. 210-235.

Dingman, S. L. 1976, The water balance in arctic and subarctic regions: annotated bibliography and preliminary assessment. U.S. Army Corps of Engineers CRREL. Spec. Rpt. 187.

French, H. M. and R. Gilbert, 1981. Periglacial phenomena near Churchill, Manitoba.Paper presented Hudson/James Bay Symposium, Guelph, Ontario, 28-30 April, 1981.

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

Gray, J. T. and R. J. E. Brown, 1979. Permafrost existence and distribution in the Chic-Chocs Mountains, Gaspésie, Quebec. Geogr. phys. Quat. 33 (3-4): 299-316.

Guymon, G.L., 1976. Summer moisture-temperature for arctic tundra. J. Irrigation and Drainage Division, ASCE 102 (IR 4), Proc. Paper 12613. pp. 403-411.

__________, 1975. Soil-moisture-temperature for Alaskan lowland. J. Irrigation and Drainage Division, ASCE 101 (IR3), Proc. Paper 11562, pp.187-199.

__________ and J. Luthin, 1974. A coupled heat and moisture transport model for arctic soils. Water Resources Res.10 (5): 995-1001.

Haag, R. W. and L. C. Bliss, 1974. Energy budget changes following surface disturbance to upland tundra. J. Applied Ecology, 11: 355-374.

Hare, F. K. and M. Thomas 1979. Climate Canada (2nd ed.) Toronto: Wiley and Sons Ltd.

Jordan, L. C., 1980. Heat and moisture exchange in a permafrost active layer, Churchill, Manitoba. Unpublished M. Sc. thesis, Department of Geography, McMaster University, Hamilton, Ont.

Judge, A. S., 1973. Thermal regime of the Mackenzie Valley. Environmental-Social Program Northern Pipelines Task Force on Northern Oil Development Report No. 73-38.

Kelley, J. and D. F. Weaver, 1969. Physical processes at the surface of the arctic tundra. Arctic 22: 425-437.

Kersten, M.S., 1949. Laboratory research for the determination of the thermal properties of soils. University of Minnesota.

Luthin, J.N. and G. L. Guymon, 1974. Soil moisture-vegetation-temperature relationships in Central Alaska. J. Hydrology 23: 233-246.

McGaw, R.W., S. I. Outcalt, and E. Ng, 1978. Thermal properties and regimes of wet tundra soils at Barrow, Alaska. Proc. Third International Conf. on Permafrost. Vol. 1. Edmonton, Alta. pp. 48-53.

Nakano, Y. and J. Brown, 1972. Mathematical modelling and validation of the thermal regimes in tundra soils, Barrow, Alaska. Arctic and Alpine Res. 4 (1): 19-38.

Ng, E. and P. Miller, 1977. Validation of a model of the effect of tundra vegetation on soil temperatures. Arctic and Alpine Res. 9 (2): 89-104.

Nicholson, F. H., 1976. Permafrost thermal amelioration test near Schefferville, Quebec. Can. J. Earth Sci. 13 (12): 1694-1705.

Oke, T. R., 1978. Boundary Layer Climates. London: Methuen 372 pp.

Outcalt, S. I., 1972. The development and application of a simple digital surface-climate simulator. J. Appl. Meteorol. 11: 629-656.

____________, C. Goodwin, G. Weller, and J. Brown, 1975. Computer simulation of the snowmelt and soil thermal regime at Barrow, Alaska. Water Resources Res. 11 (5): 709-715.

Parson, G. F., 1980. Report on the Churchill Research Centre at Churchill, Manitoba. Northern Social Research Division, Indian and Northern Affairs.

Pavlov, A. V., 1981. Thermophysical studies of natural and disturbed landscape complexes in the north of the U.S.S.R. Cold Regions Sci. Tech. 4: 129-135.

Petzold, D. E. and A. N. Rencz, 1975. The albedo of selected subarctic surfaces. Arctic and Alpine Res. 7 (4): 393-398.

Rouse, W. R., 1981a. Microclimate of low arctic tundra and forest. Presented at Fourth Can. Permafrost Conf., Calgary, Alta., 2-6 March, 1981.

____________ 1981b. The water balance of upland tundra in the Hudson Bay Lowlands - measured and modelled. Presented at the Hudson/James Bay Symposium, Guelph, Ont., 28-30 April, 1981.

____________ and K. A. Kershaw, 1973. Studies on lichen dominated systems. VI. Cdn. J. Bot. 51: 1309-1316.

____________ and R. B. Stewart, 1972. A simple model for determining evaporation from high latitude upland sites. J. Appl. Meteorol. II (7): 1063-1070.

Smith, M. W., 1975a Numerical simulation of microclimatic and active layer regimes in high arctic environment. Canada, Indian and Northern Affairs, INA Publications QS 8039-000 EE-A1, ALUR 74-75-72.

____________ 1975b. Microclimatic influence on ground temperatures and permafrost distribution, Mackenzie Delta, Northwest Territories. Can. J. Earth Sci. 12 (8): 1421-1438.

Stewart, R. and W.R. Rouse, 1976. Simple models for calculating evaporation from dry and wet tundra surfaces. Arctic and Alpine Res. 8 (3): 263-274.

Thompson, R.D. and B.D. Fahey, 1977. Energy balance and ground thermal regime studies, Boughton Island, N.W.T. Arch. fur Meteorol. Geophys. und Biokl. Ser. A. 25 (3): 251-272.

Van Wijk, W.R. and D.A. deVries, 1963. Periodic temperature variations in a homogeneous soil. Physics of the Plant Environment. Van Wijk, W.R. (ed.) New York: John Wiley and Sons, pp. 102-143.

Weller, G. and B. Holmgren, 1974. The microclimates of the Arctic tundra. J. Appl. Meteorol. 13 (8): 854-862.

Wright, R.K., 1980. The water balance of a lichen tundra underlain by permafrost. Unpublished Ph.D. thesis, Dept. of Geography, McGill University, Montreal, Canada.

APPENDIX I

CALCULATED HEAT CAPACITIES - PEAT
(MJ m^{-3} K^{-1})

	0-.10m	.10-.20m	.20-.30m	.30-.40m	.40-.50m
June 8	3.65	3.57			
10	3.65	3.57			
12	3.55	3.57			
14	3.55	3.52			
16	3.45	3.43		- FROZEN -	
18	3.40	3.40			
20	3.35	3.35			
23	3.27	3.27			
25	3.23	3.23	3.57		
28	3.18	3.03	3.57		
30	3.22	3.05	3.57		
July 2	3.28	3.05	3.57		
4	3.32	3.06	3.57		
6	3.37	3.08	3.40		
9	3.42	3.08	3.32		
11	3.47	3.10	3.30	- FROZEN -	
13	3.52	3.11	3.30		
15	3.57	3.11	3.28		
18	3.62	3.13	3.27		
20	3.65	3.13	3.25		
22	3.70	3.15	3.23		
24	3.75	3.15	3.22		
26	3.80	3.17	3.18		
27	3.85	3.18	3.18	3.57	
29	3.74	3.17	3.18	3.57	
Aug. 2	3.62	3.17	3.18	3.52	
3	3.50	3.15	3.18	3.52	
4	3.38	3.13	3.18	3.52	3.57
6	3.28	3.13	3.18	3.52	3.57
8	3.17	3.11	3.18	3.52	3.52
10	3.05	3.10	3.18	3.52	3.52
13	3.15	3.13	3.18	3.52	3.52
15	3.25	3.22	3.15	3.52	3.52
17	3.35	3.28	3.10	3.52	3.52
19	3.45	3.35	3.05	3.52	3.52
22	3.30	3.40	3.01	3.52	3.52
25	3.30	3.40	3.01	3.52	3.52
28	3.30	3.40	3.01	3.52	3.52
31	3.30	3.40	3.01	3.52	3.52

CALCULATED HEAT CAPACITIES - GRASSLAND
($MJ\ m^{-3}K^{-1}$)

	0 - .10 m	.10-.20 m	.20-.30 m	.30 -.40m	.40-.50 m	.50 - .60m
June 8	3.99	3.80	3.87			
10	3.94	3.80	3.85			
12	3.89	3.79	3.84			
14	3.82	3.79	3.82			
16	3.77	3.77	3.80	----	FROZEN	--------
18	3.72	3.75	3.79			
20	3.65	3.75	3.77			
23	3.60	3.73	3.75			
25	3.53	3.72	3.75			
28	3.48	3.70	3.74	3.90	3.90	3.90
30	3.47	3.69	3.72	3.90	3.90	3.90
July 2	3.45	3.69	3.70	3.90	3.90	3.90
4	3.41	3.66	3.70	3.90	3.90	3.90
6	3.41	3.65	3.69	3.89	3.89	3.89
9	3.42	3.64	3.67	3.89	3.89	3.89
11	3.40	3.62	3.65	3.87	3.87	3.87
13	3.38	3.60	3.64	3.87	3.87	3.87
15	3.37	3.59	3.62	3.87	3.87	3.87
18	3.35	3.57	3.62	3.85	3.85	3.85
20	3.33	3.55	3.60	3.85	3.85	3.85
22	3.33	3.55	3.59	3.85	3.85	3.85
24	3.32	3.53	3.57	3.85	3.85	3.85
26	3.30	3.52	3.55	3.85	3.85	3.85
27	3.28	3.50	3.53	3.85	3.85	3.85
29	3.35	3.56	3.57	3.85	3.85	3.85
August 2	3.42	3.65	3.62	3.87	3.87	3.87
3	3.48	3.74	3.65	3.87	3.87	3.87
4	3.55	3.80	3.70	3.89	3.89	3.89
6	3.62	3.89	3.75	3.89	3.89	3.89
8	3.69	3.79	3.79	3.90	3.90	3.90
10	3.75	4.02	3.84	3.90	3.90	3.90
13	3.69	3.97	3.87	3.92	3.92	3.92
15	3.62	3.92	3.90	3.94	3.94	3.94
17	3.57	3.85	3.94	3.94	3.94	3.94
19	3.50	3.79	3.79	3.97	3.97	3.97
22	3.43	3.74	4.02	4.02	4.02	4.02
25	3.43	3.74	4.02	4.02	4.02	4.02
28	3.43	3.74	4.02	4.02	4.02	4.02
31	3.43	3.74	4.02	4.02	4.02	4.02

CALCULATED HEAT CAPACITIES - TUNDRA
($MJ\ m^{-3}\ K^{-1}$)

	0-.10m	.10-.20m	.20-.30m	.30-.40m	.40-.50m
June 8	3.85	3.33	2.90		
10	3.74	3.27	2.90		
12	3.69	3.18	2.90		
14	3.64	3.05	2.60	2.90	
16	3.59	2.98	2.53	2.90	
18	3.52	2.90	2.46	2.71	2.90
20	3.48	2.85	2.40	2.69	2.69
23	3.43	2.76	2.33	2.65	2.65
25	3.38	2.69	2.26	2.62	2.62
28	3.33	2.61	2.19	2.19	2.19
30	3.33	2.63	2.19	2.56	2.56
July 2	3.33	2.66	2.17	2.49	2.49
4	3.33	2.67	2.17	2.46	2.46
6	3.33	2.69	2.69	2.42	2.42
9	3.33	2.73	2.17	2.40	2.40
11	3.33	2.75	2.15	2.37	2.37
13	3.33	2.76	2.15	2.33	2.33
15	3.33	2.79	2.15	2.28	2.28
18	3.33	2.80	2.15	2.26	2.26
20	3.33	2.83	2.12	2.24	2.24
22	3.33	2.86	2.12	2.12	2.12
24	3.33	2.88	2.12	2.15	2.15
26	3.33	2.90	2.10	2.10	2.10
27	3.33	2.91	2.10	2.08	2.08
29	3.33	2.91	2.10	2.08	2.08
Aug. 2	3.35	2.91	2.08	2.06	2.06
3	3.37	2.90	2.08	2.06	2.06
4	3.37	2.90	2.06	2.06	2.06
6	3.38	2.90	2.03	2.06	2.06
8	3.38	2.88	2.06	2.06	2.06
10	3.40	2.91	2.08	2.03	2.03
13	3.37	2.95	2.10	2.08	2.08
15	3.33	3.00	2.10	2.10	2.10
17	3.32	3.03	2.10	2.12	2.12
19	3.29	3.06	2.10	2.15	2.15
22	3.25	3.06	2.10	2.17	2.17
25	3.25	3.06	2.10	2.17	2.17
28	3.25	3.06	2.10	2.17	2.17
31	3.25	3.06	2.10	2.17	2.17

APPENDIX II

CALCULATED CHANGES IN SOIL HEAT STORAGE

($MJ\ m^{-2}$ measurement $period^{-1}$)

Peat - Site 1

DAY	0 - .10m	.10 - .20m
1-2	-0.7738	4.1494
2-3	2.3287	4.4752
3-4	-0.1562	4.0352
4-5	0.5716	4.1984
5-6	-1.3662	4.2455
6-7	1.0880	2.3885
7-8	1.0754	3.4399
8-9	0.8698	2.5824
9-10	-2.2126	1.2838
10-11	0.1018	1.7835
11-12	2.0898	3.7473
12-13	0.5018	2.2802
13-14	-0.0930	2.6321
14-15	-0.9874	2.1568
15-16	0.8482	-0.1047
16-17	-2.9564	-0.8308
17-18	1.1651	0.0871
18-19	0.7211	0.7899
19-20	-1.3539	-0.2848
20-21	1.0220	0.2786
21-22	-0.9398	-0.3465
22-23	2.9588	0.2898
23-24		
24-25		
25-26	-0.4413	0.2473
26-27		
27-28		
28-29	1.3689	-0.2911
29-30	-1.4399	-0.2379
30-31	1.5533	-0.2177
31-32	0.2593	0.0124
32-33	1.3514	0.5384
33-34	-2.2034	-0.3574
34-35	1.3166	-0.5084
35-36	-1.5353	0.1005
36-37	1.6500	0.6324
37-38	-2.1879	-0.9826
38-39	.0.8151	0.3468
39-40	-1.0230	-0.6256

...cont'd

...cont'd

DAY	.20 - .30m	.30 - .40m	.40 - .50m	TOTAL
1-2				3.3756
2-3				6.8039
3-4				3.8790
4-5				4.7700
5-6				2.8793
6-7				3.4765
7-8				4.5153
8-9				3.4522
9-10				-0.9288
10-11				1.8853
11-12				5.8371
12-13				2.7820
13-14				2,5391
14-15				1.1694
15-16	1.4090			2.1525
16-17	1.2195			-2.5677
17-18	1.4652			2.7174
18-19	1.6583			3.1693
19-20	1.5290			-0.1097
20-21	1.4227			2.7233
21-22	1.4019			0.1156
22-23	1.5299			4.7785
23-24				
24-25				
25-26	1.4733			1.2793
26-27				
27-28				
28-29	1.7595			2.8373
29-30	0.0111	3.1976		1.5309
30-31	-0.1797	2.6080		3.7639
31-32	0.0620	2.9917		3.3254
32-33	0.1336	2.9054		4.9288
33-34	-0.0331	2.9406		0.3467
34-35	-0.5410			0.2672
35-36	0.4956			
36-37	0.2980	0.3326	3.1430	6.0560
37-38	-0.6712	2.1152		-1.7265
38-39	0.2995	3.4827		4.9441
39-40	-0.4485	2.3704		0.2733

Peat - Site 2

DAY	0 - 10 m	.10 - .20 m	.20 - .30 m	.30 - .40 m
1-2	-1.6790	1.4293		
2-3	4.5844	1.5685		
3-4	-2.7087	1.3656		
4-5	0.7668	1.9534		
5-6				
6-7				
7-8	1.1524	1.8287		
8-9	1.1936	2.0006		
9-10	-1.1693	1.0523		
10-11	0.9985	4.5230		
11-12	0.5410	4.1897		
12-13	-0.4756	4.7143		
13-14	1.2417	4.9577		
14-15	-0.8560	3.8047		
15-16	-0.3694	-0.1663	1.9993	
16-17	-2.2624	0.6169	1.9933	
17-18	0.4752	0.0746	2.6573	
18-19	1.3959	0.7899	2.0220	
19-20	-0.7711	-0.1941	2.1422	
20-21	1.4746	0.1033	1.9982	
21-22	-1.2358	-0.3906	2.1385	
22-23	1.6875	0.6332	2.2723	
23-24				
24-25	-1.4130	-0.5491	-0.1574	2.0931
25-26				
26-27				
27-28	-1.5750	-0.2646	0.0827	0.1038
28-29	1.0496	0.3161	0.1161	0.0370
29-30	-1.3153	-0.4163	-0.1526	-0.0510
30-31	0.2029	0.0031	0.0525	0.0880
31-32	-0.3752	-0.1023	-0.2401	-0.1408
32-33	1.8396	0.1628	0.2576	0.2482
33-34	-1.8818	0.3735	0.2032	0.0211
34-35	0.9682	-0.1542	-0.1287	0.3854
35-36	-0.7659	-0.2881	0.1571	-0.0176
36-37	1.4652	0.5712	0.1204	0.0493
37-38	-2.5344	0.9826	-0.7495	-0.7990
38-39				
39-40				

....cont'd

....cont'd

DAY	.40 - 50 m	.50 - .60 m	.60 - .70 m	TOTAL
1-2				-0.2497
2-3				6.1529
3-4				-1.3431
4-5				2.7202
5-6				
6-7				...
7-8				2.9811
8-9				3.1942
9-10				-0.1170
10-11				5.5215
11-12				4.7307
12-13				4.2387
13-14				6.1994
14-15				2.9487
15-16				1.4636
16-17				0.3478
17-18				3.2071
18-19				4.2078
19-20				1.1770
20-21				3.5761
21-22				0.5121
22-23				4.5930
23-24				
24-25				
25-26				
26-27				
27-28	2.6955			1.0424
28-29	2.6260			4.1448
29-30	2.6117			0.6765
30-31	2.6682			3.0147
31-32	2.6206			1.7623
32-33	2.7843			5.2925
33-34	-0.0933	10.4332		9.0559
34-35	0.6846	11.0615		12.8208
35-36	-0.2306	-0.1848	10.4701	9.1402
36-37	-0.0334	-0.0669	10.4033	12.5091
37-38	-0.5157	6.7149		3.0989
38-39				
39-40				

Grassland - Site 3

DAY	0 - .10 m	.10 - .20 m	.20 - .30 m
1-2	-1.4324	2.1541	
2-3	2.9195	2.5759	
3-4	-1.2759	2.1022	
4-5	-0.1757	1.8814	
5-6	4.6409	2.8002	
6-7	-4.8323	1.6424	
7-8	1.1315	2.7482	
8-9	1.2744	3.6061	
9-10	-0.9919	1.4597	
10-11	0.5672	2.6440	
11-12	0.3834	2.3034	
12-13	1.1420	2.7388	
13-14	-0.5251	0.3697	6.5065
14-15	0.2523	0.6607	7.2548
15-16	-0.8276	-0.9937	6.5813
16-17			
17-18			
18-19	1.6378	1.1273	0.5521
19-20	-0.4456	0.1321	0.0652
20-21	-0.0298	-0.3723	-0.0504
21-22	-0.3730	-0.1207	0.0539
22-23	0.5013	0.3530	0.0964
23-24			
24-25			
25-26	-0.6298	-0.6978	-0.4944
26-27			
27-28			
28-29	0.3905	0.1406	0.1351
29-30	-0.8435	-0.5524	-0.2906
30-31	0.4059	0.1402	0.1649
31-32	-1.0350	-0.4342	-0.1459
32-33	0.9999	0.3970	0.1451
33-34	-0.0072	0.7017	0.4563
34-35	-0.8675	-1.1127	-0.6797
35-36	-0.1750	-0.0076	0.0754
36-37	0.4425	0.0486	-0.2794
37-38	-0.6346	-0.3516	-0.1427
38-39	-0.2710	-0.4114	-0.0402
39-40			

....cont'd

...cont'd

DAY	.30 - .40m	.40 - .50m	.50 - .60m	DAILY TOTAL
1-2				0.7217
2-3				5.4954
3-4				0.8263
4-5				1.7057
5-6				7.4411
6-7				-3.1899
7-8				3.8797
8-9				4.8805
9-10				0.4678
10-11				3.2112
11-12				2.6868
12-13				3.8808
13-14				6.3511
14-15				8.1678
15-16				4.7600
16-17				
17-18				
18-19	9.3163			12.6335
19-20	9.1654			8.9171
20-21	0.1020	1.9896		1.6391
21-22	0.1463	2.0782		1.7847
22-23	1.6739			2.6246
23-24				
24-25				
25-26	2.5325			0.7105
26-27				
27-28				
28-29	0.1595	2.0423		2.8680
29-30	-0.1303	1.8886		0.0718
30-31	0.1443	2.0542		2.9095
31-32	-0.0468	1.9547		0.2928
32-33	0.0353	2.0016		3.5789
33-34	0.2344	2.0491		3.4343
34-35	-0.2561	-0.0256	5.6184	2.6768
35-36	-0.1390	-0.0695	5.5238	5.2081
36-37	-0.4482	5.8574		5.6209
37-38	0.2151	5.7750		4.8612
38-39	-0.0382	0.0080	5.6625	4.9097
39-40				

Grassland - Site 4

DAY	0 - .10m	.10 - .20m	.20 - .30m	.30 - .40m	.40 - .50m
1-2	1.3636	2.1503			
2-3	2.4389	2.6035			
3-4	-0.9375	2.1999			
4-5	1.0467	2.6510			
5-6	-1.5419	1.7019			
6-7	0.8816	2.9520			
7-8	0.3504	2.9858			
8-9	2.7720	2.5047			
9-10	-1.8250	2.5600			
10-11	-0.0905	2.8398			
11-12	1.4782	2.2812			
12-13	-0.2277	2.7978			
13-14	1.3197	0.7613	6.8580		
14-15	-0.9275	0.7446	8.0279		
15-16	-1.1902	-1.7727	5.5372		
16-17	-1.1356	-0.4344	6.9901		
17-18	0.2501	-0.2232	-0.2402	3.7994	
18-19	1.6109	0.9226	0.6118	4.3122	
19-20					
20-21					
21-22	-0.1765	-0.2024	-0.0090	4.1505	
22-23					
23-24					
24-25	-0.4756	-0.1540	-0.0141	0.0231	5.4604
25-26	-0.7203	-0.6550	-0.2517	-0.0212	5.6241
26-27					
27-28					
28-29	0.8485	0.5016	0.2424	0.1323	5.6737
29-30	-0.8000	-0.4746	-0.1444	-0.0156	0.0019
30-31	-0.1550	-0.2198	-0.2653	-0.1677	-0.0429
31-32	-0.6038	0.1246	0.0154	0.0604	0.0371
32-33	1.4059	0.2025	0.2380	0.0470	-0.0098
33-34	-1.0245	0.3842	0.2886	0.1162	0.0670
34-35	-0.2999	-0.7508	-0.4078	-0.0039	-0.0394
35-36	-0.4795	-0.1668	-0.0794	-0.1906	-0.1608
36-37	0.8644	0.3403	0.1869	0.1387	0.1146
37-38	-1.2348	-0.7704	-0.6070	-0.4764	-0.4925
38-39	0.0892	-0.2169	0.0583	0.2332	0.3558
39-40	-0.3430	-0.0374	-0.1849	-0.1668	-0.0663

....cont'd

cont'd..

DAY	50-60 m	60 - 70 m	Daily Total
1-2			3.5149
2-3			5.0424
3-4			1.2624
4-5			3.6977
5-6			0.1600
6-7			3.8336
7-8			3.3662
8-9			5.2767
9-10			0.7350
10-11			2.7493
11-12			3.7594
12-13			2.5701
13-14			8.9390
14-15			7.8450
15-16			6.1197
16-17			5.4201
17-18			3.5861
18-19			7.4575
19-20			
20-21			
21-22			3.7626
22-23			
23-24			
24-25			4.8398
25-26			3.9759
26-27			
27-28			
28-29			7.3985
29-30	13.8720		12.4393
30-31	13.7961		12.9454
31-32	0.0683	3.1393	2.8105
32-33	-0.0098	3.0279	4.9017
33-34	0.0433	3.1537	5.0775
34-35	0.0276	3.0474	1.6520
35-36	-0.1112	3.0055	1.8172
36-37	0.1407	3.2037	4.9893
37-38	3.5263		-0.0548
38-39	3.5494		4.0690
39-40	-0.0322	3.0911	2.2605

Tundra - Site 5

DAY	0 - .10 m	.10 - .20 m	.20 - .30 m	.30 - .40 m
1-2	-1.0742	2.3259		
2-3	2.5544	2.9865		
3-4	-1.3616	3.1147		
4-5	1.4487	2.4841		
5-6	1.7053	2.9587		
6-7	-2.2810	-0.6525	2.1281	
7-8	1.0510	0.2537		
8-9	1.9482	0.7590		
9-10	-2.4843	0.3901	2.3799	
10-11	-0.9391	-0.0548	2.3957	
11-12	-0.3596	0.4050	2.6935	
12-13	1.0290	0.4868	0.0738	-0.0971
13-14	-0.0133	-0.0961	0.0043	0.0898
14-15	0.1598	0.4869	0.5151	0.8487
15-16	-0.5528	-0.3931	-0.5013	-0.4200
16-17				
17-18	-0.1631	-0.6596	-0.2946	-0.0977
18-19	-0.1732	9.3257	0.3291	0.3448
19-20				
20-21	0.6627	-0.0481	0.1230	0.1322
21-22				
22-23	-0.7959	0.3596	0.3003	0.2919
23-24				
24-25	2.2244	0.1659	-0.0662	-0.1134
25-26	-2.9304	-0.8410	-0.4137	-0.3630
26-27	2.2814	0.7509	0.3422	0.2884
27-28	-2.0928	-0.9164	0.4274	-0.3863
28-29	0.4415	0.3828	0.3203	0.3657
29-30	0.7098	0.4176	0.0863	0.0824
30-31				
31-32				
32-33	1.5367	0.5587	0.2276	0.1279
33-34	-1.7420	-0.0797	0.2392	0.2667
34-35	1.2383	-0.0420	-0.3150	-0.2576
35-36	-1.7290	-0.5728	-0.0682	-0.0473
36-37	0.9003	0.6028	0.1985	0.1747
37-38	-1.7290	-1.3709	-0.7539	-0.7421
38-39	0.9003	0.4379	0.3287	0.3613
39-40	-1.2383	-0.1010	-0.0189	-0.0195

...cont'd

...cont'd

DAY	40 - .50m	.50 - 60m	60 - .70m	.70 - .80m	.80 - .90m
1-2					
2-3					
3-4					
4-5					
5-6					
6-7					
7-8					
8-9					
9-10					
10-11					
11-12					
12-13	4.5672				
13-14	4.7954				
14-15	5.4547				
15-16	-0.2604	-0.0876	9.5417		
16-17					
17-18	0.0233	0.0746	0.1398	9.5251	
18-19	0.3120	0.2638	0.2269	0.0114	0.0091
19-20					
20-21	0.1568	0.3024	0.4435	0.3248	0.3270
21-22					
22-23	0.3129	0.1386	0.1050	0.0651	0.0756
23-24					
24-25	-0.1300	-0.0718			
25-26	-0.3411	-0.3474	-0.3650	-0.3370	-0.3058
26-27	0.3008	0.3533	0.3852	0.3945	0.3842
27-28	-0.4305	-0.4305	-0.4192	-0.4213	-0.4130
28-29	0.4563	0.4676	0.4542	0.4563	0.4470
29-30	0.0360	0.0082	0.0237	0.0330	0.0299
30-31					
31-32					
32-33	0.0822	0.0250	-0.0198	-0.0270	-0.0322
33-34	0.2856	0.2436	0.1806	0.1775	0.1754
34-35	-0.2258	-0.1283	-0.0127	0.0085	-0.0191
35-36	-0.0709	-0.0838	-0.0742	-0.0645	-0.0623
36-37	0.1421	0.0911	0.0738	-0.5609	0.0608
37-38	-0.6825	-0.5968	-0.5761	-0.5609	-0.5295
38-39	0.3526	0.3168	0.3320	0.3591	0.3656
39-40	0.0054	0.0206	-0.0011	-0.0076	0.0109

....cont'd

...cont'd

DAY	90 - 100 m	100 - 110 m	110 - 120 m	120 - 130 m	TOTAL
1-2					1.2517
2-3					5.5409
3-4					1.7531
4-5					3.9328
5-6					4.6640
6-7					-0.8054
7-8					1.3047
8-9					2.7072
9-10					-0.4945
10-11					1.4018
11-12					2.7389
12-13					6.0597
13-14					4.7801
14-15					7.4652
15-16					7.3265
16-17					
17-18					8.5478
18-19	-0.0046	1.3236			1.7774
19-20					
20-21	0.3069	1.7164			4.4476
21-22					
22-23	0.0714	1.5096			2.7937
23-24					
24-25					2.0089
25-26	-0.3037	1.1521			-5.0486
26-27	0.3749	1.8089			5.6113
27-28	-0.4151	1.0158			-4.4819
28-29	0.4522	0.4511	2.8201		7.5151
29-30	0.0319	-0.0051	2.3371		3.7908
30-31					
31-32					
32-33	-0.0239	2.3719			4.8271
33-34	0.1638	2.5356			2.4463
34-35	0.0265	0.0307	0.0329	2.4180	2.7926
35-36	-0.0591	-0.0473	-0.0312	-2.3682	-0.5424
36-37	0.0673	0.0694	0.0618	2.4527	4.3344
37-38	-0.5034	-0.4785	1.9384		-6.5852
38-39	2.7489				6.5032
39-40	2.4082				1.0587

****

Tundra - Site 6

DAY	0 - .10m	.10 - .20m	.20 - .30m	.30 - .40m
1-2	-1.2474	2.2891		
2-3	2.3712	2.8091		
3-4	-0.5535	2.2084		
4-5	1.1029	2.5451		
5-6				
6-7				
7-8	-0.3097	0.1767	4.4798	
8-9	2.2295	0.7342	3.9596	
9-10				
10-11				
11-12	-0.1265	0.2025	0.1697	0.2074
12-13	3.1036	0.2580	0.2832	0.2291
13-14	-1.6750	-0.1629	-0.4438	-0.0576
14-15	-0.4562	0.9630	1.2939	1.2741
15-16	0.4529	-0.7671	-0.6163	-0.6552
16-17				
17-18				
18-19	-0.1632	0.7728	0.3623	2.0781
19-20 20-21	0.8225	0.0651	0.0742	0.0023
21-22	-0.8292	-0.3861	-0.0816	0.0329
22-23				
23-24				
24-25				
25-26				
26-27				
27-28				
28-29	0.1078	0.0667	0.0946	0.0360
29-30	9.8044	0.4176	0.0728	0.1061
30-31	-0.2028	-0.0232	0.0680	0.0597
31-32	0.0850	-0.2736	-0.1604	-0.1533
32-33	0.9807	0.0960	0.1411	0.1362
33-34	-1.5518	0.2360	0.2100	0.1838
34-35	0.3254	-0.4370	-0.3224	-0.2555
35-36	-1.0824	-0.1303	0.0021	0.0140
36-37	0.9555	0.1989	0.0788	0.0195
37-38	-0.9848	-1.0557	-0.7560	-0.7671
38-39	1.0303	0.8170	0.4263	0.5327
39-40				

...cont'd

...cont'd

DAY	.40 - .50m	.50 - .60m	.60 - .70m	.70 - .80m	.80 - .90m
1-2					
2-3					
3-4					
4-5					
5-6					
6-7					
7-8					
8-9					
9-10					
10-11					
11-12	0.1715	1.8692			
12-13	0.1656	2.1055			
13-14	1.2797				
14-15	3.1598				
15-16	1.3081				
16-17					
17-18					
18-19					
19-20					
20-21	0.1366	0.2240	1.7610		
21-22	0.0498	0.0329	1.6037		
22-23					
23-24					
24-25					
25-26					
26-27					
27-28					
28-29	0.0484	0.1164	0.1112	0.1051	0.1267
29-30	0.0824	0.0319	-0.0082	-0.0361	-0.0412
30-31	0.0567	0.0422	0.0670	0.0968	0.0876
31-32	-0.1157	-0.0670	-0.0325	0.0010	0.0233
32-33	0.0759	0.0426	0.0333	0.0208	0.0135
33-34	0.1565	0.1176	0.0840	0.0714	0.0578
34-35	-0.1738	-0.1113	-0.0657	-0.0413	-0.0148
35-36	0.0054	-0.0183	-0.0495	-0.0699	-0.0774
36-37	0.0033	0.0011	0.0043	0.0174	0.0217
37-38	-0.7497	-0.6868	-0.6098	-0.5794	-0.5317
38-39	0.5968	0.5978	0.5609	0.5197	0.4698
39-40					

...cont'd

...cont'd

DAY	.90 - 100 m	100 - 110m	110-120m	120-130m	TOTAL
1-2					1.0417
2-3					5.1803
3-4					1.6549
4-5					3.6480
5-6					
6-7					
7-8					4.3468
8-9					6.9233
9-10					
10-11					
11-12					2.4938
12-13					6.1450
13-14					-1.0596
14-15					6.2346
15-16					-0.2776
16-17					
17-18					
18-19					3.0500
19-20					
20-21					3.0857
21-22					0.4224
22-23					
23-24					
24-25					
25-26					
26-27					
27-28					
28-29	0.1401	4.9265			5.8803
29-30	-0.0721	4.6947			6.0565
30-31	0.1082	4.8750			5.2352
31-32	0.0132	0.0081	7.1686		6.4967
32-33	0.0104	0.0166	7.1917		8.7588
33-34	0.0525	0.0525	0.0536	7.0902	6.8141
34-35	0.0021	0.0117	0.0170	7.1584	6.0928
35-36	-0.0656	-0.0462	3.5561		1.6380
36-37	0.0054	-0.0152	3.5494		4.8401
37-38	-0.4676	-0.4318	3.1838		-4.4366
38-39	0.4242	0.4058	3.9758		10.3571
39-40					

****

...cont'd

DAY	90 - 100 m	100 - 110m	110-120m	120-130m	TOTAL
1-2					1.0417
2-3					5.1803
3-4					1.6549
4-5					3.6480
5-6					
6-7					
7-8					4.3468
8-9					6.9233
9-10					
10-11					
11-12					2.4938
12-13					6.1450
13-14					-1.0596
14-15					6.2346
15-16					-0.2776
16-17					
17-18					
18-19					3.0500
19-20					
20-21					3.0857
21-22					0.4224
22-23					
23-24					
24-25					
25-26					
26-27					
27-28					
28-29	0.1401	4.9265			5.8803
29-30	-0.0721	4.6947			6.0565
30-31	0.1082	4.8750			5.2352
31-32	0.0132	0.0081	7.1686		6.4967
32-33	0.0104	0.0166	7.1917		8.7588
33-34	0.0525	0.0525	0.0536	7.0902	6.8141
34-35	0.0021	0.0117	0.0170	7.1584	6.0928
35-36	-0.0656	-0.0462	3.5561		1.6380
36-37	0.0054	-0.0152	3.5494		4.8401
37-38	-0.4676	-0.4318	3.1838		-4.4366
38-39	0.4242	0.4058	3.9758		10.3571
39-40					

****