Nordic Hydrology, 27, 1996, 129-142

Effects of Hydrology on the Thermal Conditions of the Active Layer

Paper presented at the 10th Northern Res. Basin Symposium (Svalbard, Norway - 28 Aug./3 Sept. 1994)

Ming-ko Woo and Zhaojun Xia
McMaster University, Hamilton, Ont. L8S 4K1 Canada

Ground thawing and soil warming are governed by ground heat flux, soil thermal properties and the ice content of the soil, all of which are directly or indirectly influenced by the soil moisture status. Ground temperature and moisture were measured at two Arctic sites: a wetland site of frequent saturation and an adjacent pebbly loam site which was much drier. Both thermal conductivity and heat capacity were strongly affected by the ice and water contents. At both sites, about half of the ground heat flux was consumed by latent heat for ground thawing and a large fraction of heat was also conducted from the seasonally thawed zone into the permafrost, leaving only a small amount of heat to warm the active layer. The wetland soil had a shallower maximum depth of thaw than the drier site and this was due to the large ice content in its active layer. Our results demonstrate the ground thaw response to the thermal properties of the soil and its ice content, both of which are influenced by the hydrological conditions of the active layer.

Introduction

In continuous permafrost areas, seasonal thawing of the active layer makes it possible for many hydrological processes to occur, including groundwater storage and flow, infiltration of rainwater and evaporation of the soil moisture (Woo 1988). The thickness of this seasonally thawed zone and the duration of thaw are therefore important to the hydrology of the active layer. Conversely, hydrological conditions affect the ground thaw. Heat input to the active layer is supplied by the energy flux at the surface. This heat is used for ground thaw and to warm the active layer and the permafrost below. Heat transfer and the melting of ground ice are influenced strongly by the moisture status of the soil, both in its thawed and frozen states. The purpose of this paper is to assess quantitatively for two typical arctic soils, the fraction of the ground heat consumed by thawing and by soil warming. The results will elucidate the role of hydrology in influencing the active layer thaw.

Computation of Heat Fluxes in the Soil

Active layer thaw usually begins after snow melt, following the removal of a 0°C boundary condition imposed by the snow upon the ground surface. We consider the one-dimensional heat balance of the active layer for the period between the snow disappearance and the freeze-back

$$Q_g = Q_l + Q_s + Q_p \tag{1}$$

where Q_g is the ground heat flux into the soil, Q_l is the latent heat used to thaw the ice in the soil, Q_s is the heat that warms the active layer and Q_p is the heat conducted out of the active layer into the permafrost below.

Q_g consists of heat conduction and the convection of heat by infiltrating rainwater

$$Q_g = K \left.\frac{dT}{dz}\right|_{\mathrm{surface}} + c_w \Delta T \frac{dF}{dt} \tag{2}$$

Here, K is the thermal conductivity of the soil, dT/dz is the temperature gradient at the surface, c_w is volumetric heat capacity of water, ΔT is the difference in temperature between rainwater and the soil and dF/dt is the rate of rainwater infiltration. In all computations, $dt \approx \Delta t$ is taken to be one day.

Downward conduction of heat to the permafrost is

$$Q_p = K \left.\frac{dT}{dz}\right|_{\mathrm{bottom}} \tag{3}$$

Here, dT/dz is the temperature gradient at the bottom of the active layer. The heat that warms the active layer is calculated by

$$Q_s = C \frac{dT}{dt} Z \tag{4}$$

with C being the heat capacity of the soil and dT/dt is the daily temperature change of the active layer which has a thickness of Z.

If the fractional ice content in the soil (f_{ice}) and the rate of ground thaw (dh/dt) are known, Q_l can be computed by

$$Q_l = \rho \lambda f_{ice} \frac{dh}{dt} \tag{5}$$

where ρ is the density of ice and λ is the latent heat of fusion. In this study, Q_l is obtained as the residual term of Eq. (1), but Eq. (5) provides a check on the calculations.

In the above equations, the bulk thermal conductivity for a soil layer is estimated by (Farouki 1981)

$$K = \prod_j k_j^{f_j} \tag{6}$$

and the heat capacity is determined by (de Vries 1963)

$$C = \sum_j c_j f_j \tag{7}$$

with k, c and f being the thermal conductivity, heat capacity and fractional contents of various soil constituents j, including the minerals, organic, air, water and ice components. The fractions of ice, water or air that occupy the soil pores vary during the season so that the thermal conductivity and heat capacity are strongly influenced by the freeze-thaw and wet-dry transitions of the active layer.

Study Site

Two sites in the continuous permafrost area of Arctic Canada provided the ground temperature and hydrological data required by this study. These sites are located at Resolute, Cornwallis Island, Northwest Territories (74°55'N, 94°51'W) where the results of a previous hydrological investigation have been reported (Woo and Marsh 1990). A relatively dry site is underlain by a mixture of pebbles and loam, designated as polar desert soil (Cruickshank 1971). The other site, 40 m away, is situated in a fen, with a surface layer of living moss and peaty materials overlying silty clay. The mineral, organic and ice contents of the two soils were determined by digging frozen samples from the sites at various depths. Over 60 samples were collected and oven-dried in Resolute to obtain the ice content. The dried samples were then burnt at McMaster University to determine the organic content. Table 1 summarizes the soil constituents for various depths in the active layer.

The study period spanned from June 9 to August 24, 1988 but for this study, only the thawing period (July 2 to August 16) is considered. At both sites, temperature of the active layer was measured using thermistors. For the fen, temperatures were read daily at depths of 0.02, 0.1, 0.2 and 0.3 m. For the polar desert soil, the depths were 0.02, 0.1, 0.25 and 0.5 m. The unfrozen soil moisture content of the soils was measured using a time-domain reflectometer (TDR) with probes set for depths of 0.1, 0.2 and 0.3 m at the fen site and 0.1, 0.25 and 0.5 m at the polar desert site. Depth to the

Table 1 – a) Volumetric composition of soils at experimental sites and b) Thermal properties of materials used for calculations (after de Vries 1963 and Farouki 1981)

a)	Depth (m)	Mineral Content	Organic Content	Porosity	Pre-melt moisture (ice and water)
Polar Desert					
pebbly loam	0.05-0.15	0.47	0.10	0.43	0.31
pebbly loam	0.15-0.45	0.56	0.08	0.37	0.34
pebbly loam	0.45-0.55				
	frozen	0.30	0.08	0.62	0.60
	thawed	0.38	0.10	0.52	-
Fen					
peaty clay	0.02-0.10	0.14	0.06	0.80	0.66
silty clay	0.1-0.2	0.23	0.10	0.67	0.63
silty clay	0.2-0.3	0.36	0.09	0.55	0.55

b)	Thermal conductivity W/(m°C)	Heat capacity MJ/(m^{3}C)
mineral	2.93	1.926
organic	0.25	2.508
ice	2.20	1.900
water	0.57	4.180
air	0.025	0.00125

frost table was measured by hammering a steel rod into the active layer until frost was encountered. This method is prone to an error of about 2 cm at the polar desert, but reached 4 cm at the fen site where the saturated ground was easily compacted by the weight of the observer.

Hydrological Conditions

The wetland environment of the fen site produced abundant seasonal ground ice in its frozen active layer. Pore ice was common within the top 0.1 m and reticulate ice, which filled horizontal and vertical cracks, was also found below that depth (Fig. 1). The volumetric ice content ranged from 55 to 66% (Table 1). The polar desert soil had mostly pore ice at the top 0.35 m and its ice content was about 30%. Below that depth, layers of segregated ice were encountered, raising the ice content to 60% in the loamy soil (Fig. 1). This high ice content was probably fed by water at the lower portion of the active layer which was saturated in the late summer of the previous year. After thawing, soil samples from this depth yielded an average porosity of 0.52, probably because of the loss of voids after ice melt.

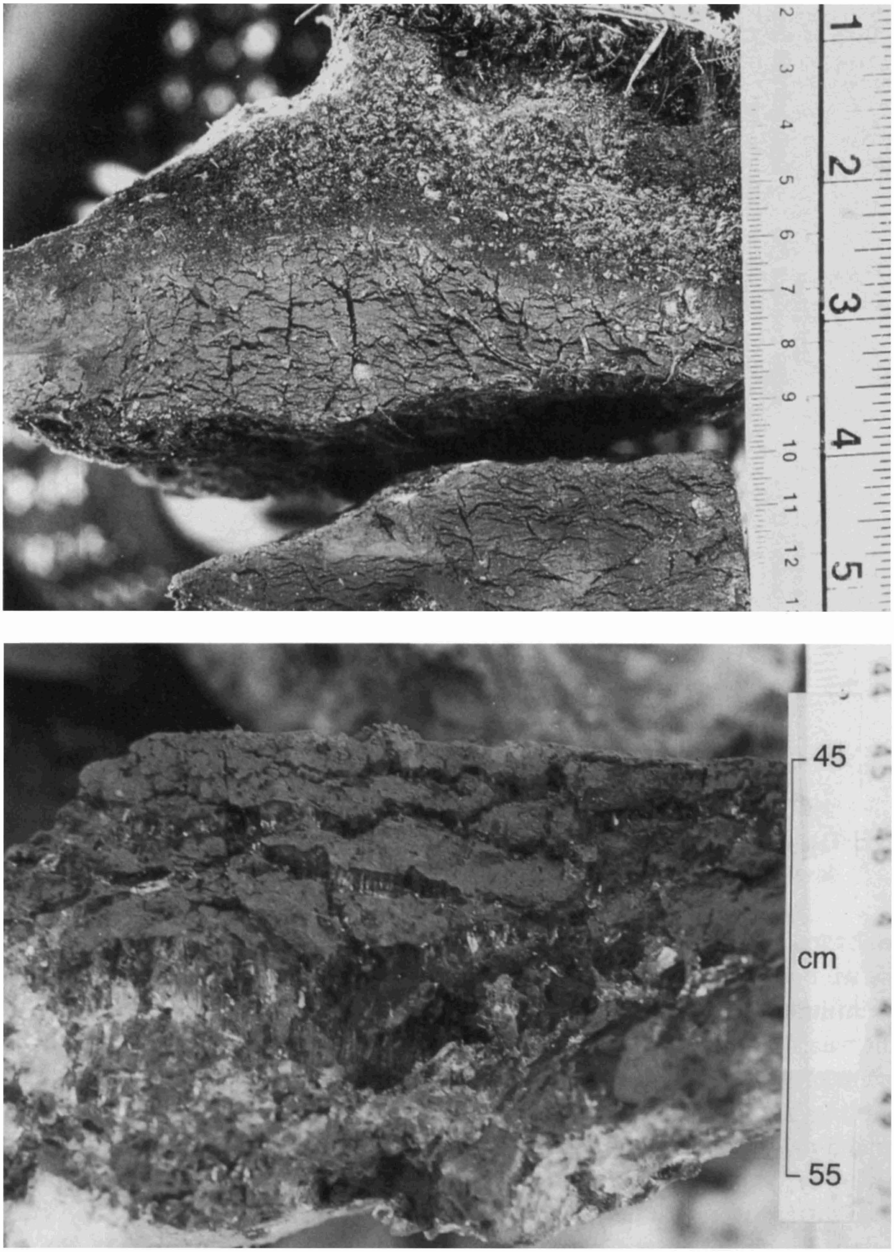

Fig. 1. Seasonal ground ice in frozen soils. a) fen soil consisting of 1 cm of living moss on top of 6 cm of peaty clay with many plant roots and abundent pore ice. Below that is clay with reticulate ice which filled vertical and horizontal shrinkage cracks.
b) polar desert soil at 45-55 cm below the surface showing layers of segregated ice, some reaching > 1 cm thickness, formed in loamy material.

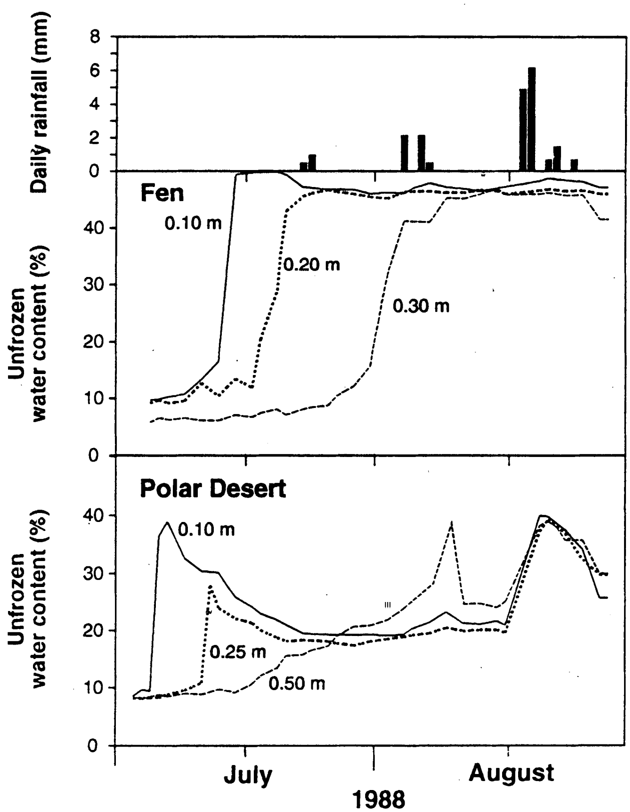

Fig. 2. Daily variation of unfrozen water content at three depths in the fen and the polar desert soils. Daily rainfall is also shown.

Snowmelt in 1988 occurred between 22 June and 1 July during which period the soils were warmed partly by the heat released through the refreezing of meltwater that infiltrated the frozen soil (Woo and Marsh 1990). At the beginning of thaw, TDR measurements indicated the presence of unfrozen water in the soil. The water content of various soil layers increased sharply during the passage of the thawing front (Fig. 2). For the fen soil, a high water table was sustained throughout the summer and the water content of the thawed soil was maintained at around 0.5. When thawing began at the polar desert soil, its near surface zone was rapidly saturated by the residual snow meltwater, to reach a maximum moisture content of 0.4. As drying continued in July, the water content decreased except near the base of the active layer where thawing of the segregated ice gradually released unfrozen water to the soil (Fig. 2). After thawing, lateral drainage of the water near 0.5 m reduced the water content at depth. Rainfall in August raised the moisture level of the entire active layer at the polar desert site. Rain had little effect on the water table position in the fen as this site was saturated continuously.

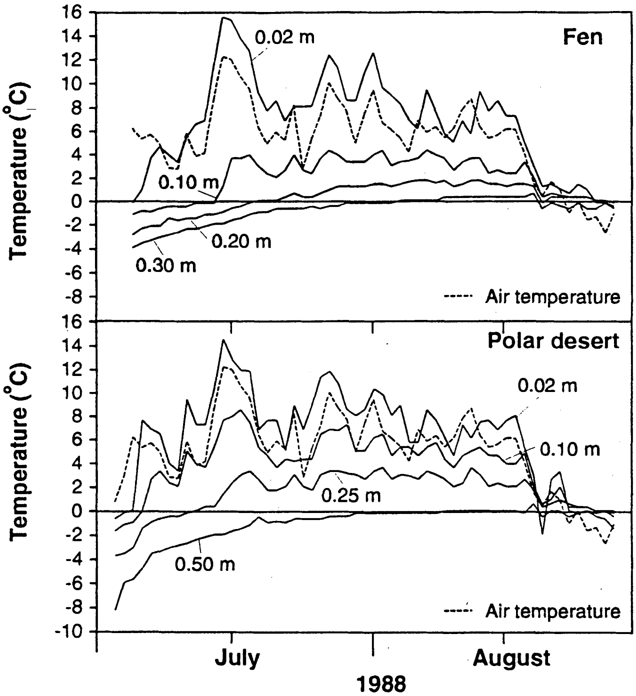

Fig. 3. Air temperature (dashed lines) and ground temperatures at various depths of the active layer, fen and polar desert soils.

Active Layer Temperature

Temperature observations spanned the entire thawing season, during which the near-surface temperatures reflected the air temperature fluctuations (Fig. 3). The surface temperature at the fen showed greater variations than the polar desert soil. At 0.1 m and below, however, the polar desert soil was always warmer than the soil of the fen. The temperature profiles showed larger temperature gradients at the surface 0.1 m, being two to three times larger than the gradient below (Table 2).

Table 2 – Average temperature gradients, July 2-August 16, 1988

	Depth (m)	Gradient (°C/m)
Polar desert site	0.02 - 0.10	37.1
	0.10 - 0.25	18.4
	0.25 - 0.50	11.6
Fen site	0.02 - 0.10	67.6
	0.10 - 0.20	20.0
	0.20 - 0.30	11.7

While ground thaw began on the same day (2 July) at both sites, the maximum thaw depth in the polar desert soil reached 0.55 m, but was only slightly deeper than 0.3 m in the fen. Air temperature fell to 0°C on 17 August and the ground temperatures plunged accordingly as freeze-back of the active layer began.

Thermal Conductivity and Heat Capacity

The fact that $k_{mineral} > k_{ice} > k_{water} > k_{organic} > k_{air}$ (where k is the thermal conductivity, Table 1) significantly affects the bulk thermal conductivity of the soil through the fractional composition of its various constituents (Eq. (6)). The fractions of ice, water and air in the soil change with freeze-thaw and wet-dry conditions to render the thermal conductivity of the active layer highly variable during the summer.

The surface peat layer at the fen site had lower thermal conductivity than its underlying silty clay because of its low mineral content but high porosity. The entire soil column has high ice contents when frozen. Upon thawing, water replaced ice and reduced the conductivity sharply (Fig. 4). As this site was saturated throughout the summer, the water fraction was little perturbed after thaw and the thermal conductivity remained relatively unchanged. The drier polar desert site experienced sur-

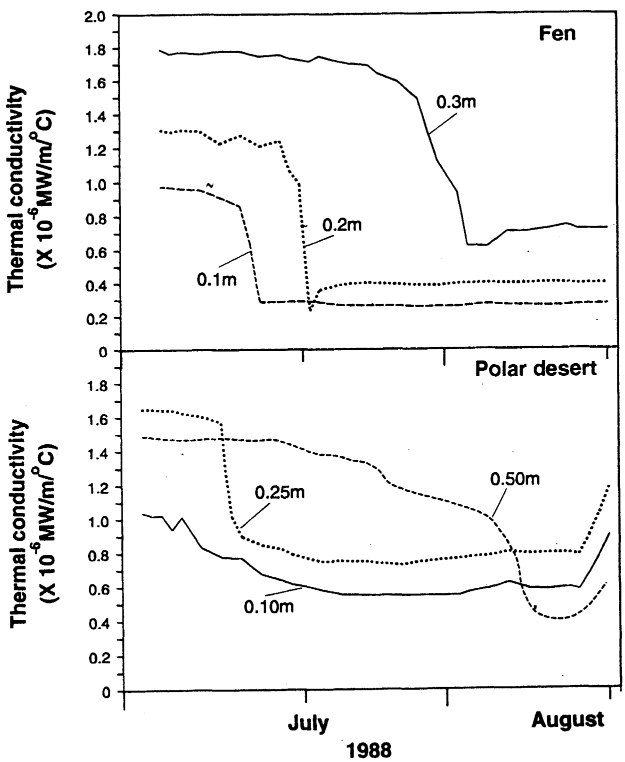

Fig. 4. Daily variations in thermal conductivity at various depths, fen and polar desert soils.

face thaw soon after snowmelt when the infiltrated meltwater filled many of the voids. As drying followed, air began to replace water and the conductivity decreased at the surface layer. Continual decline in thermal conductivity at 0.5 m was due to gradual replacement of the ground ice by unfrozen water, until all the ice thawed around August 6. Conductivity at all levels rose during the rain event of mid-August, as water replaced part of the air in the voids of the polar desert soil.

The bulk heat capacity of the soil is computed using Eq. (7). The fen soil showed a rapid increase in its bulk heat capacity as it thawed. This is attributed to the fact that the heat capacity of water far exceeds that of ice (Table 1). This phenomenon caused the bulk heat capacity of the fen soil to increase rapidly upon thawing. At this wetland site, saturation conditions both with ice when frozen, and with water when thawed, gave rise to a simple switch from low to high heat capacity during the summer (Fig. 5). Where the soil was not constantly saturated at the polar desert site, the surface zone witnessed an increase in heat capacity when it was saturated by snowmelt water. The summer decline in heat capacity was related to drying. Again, the replenishment of soil moisture by rain water raised the heat capacity. For this soil, wetting and drying rather than phase change, have more profound influence upon the heat capacity changes.

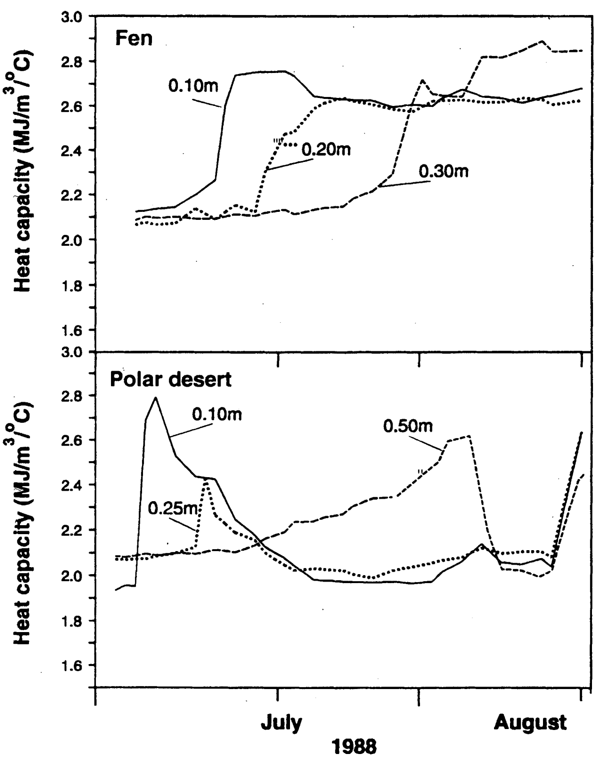

Fig. 5. Daily variations in heat capacity at various depths, fen and polar desert soils.

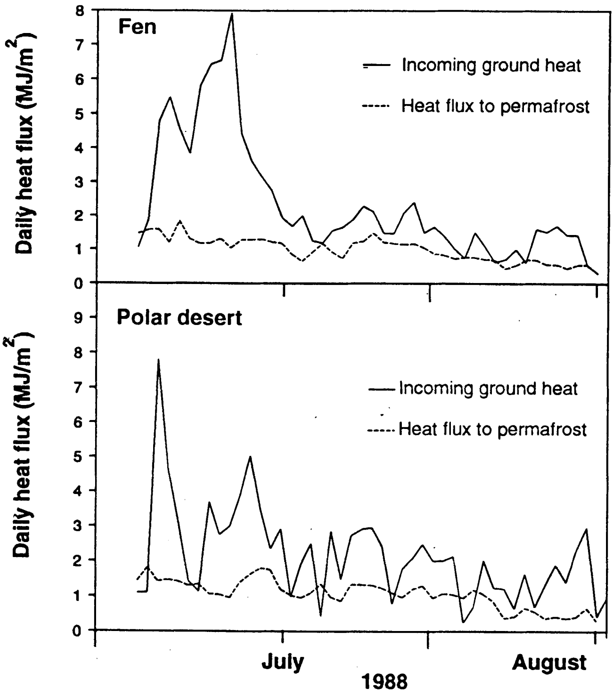

Fig. 6. Daily incoming ground heat flux into the active layer and outgoing heat flux from the active layer to the permafrost below, fen and polar desert soils.

Ground Heat Balance

Heat flux into the soil is calculated by Eq. (2), using the near surface temperatures to obtain the temperature gradients. During the study period, the mean daily near surface temperature gradient was positive so that the heat flux was directed downward, reversing direction only when freeze-back occurred in late August. Heat flux was the highest soon after snowmelt, at a time when both the thermal conductivity and the temperature of the near surface layer were large. Day to day fluctuations were evident (Fig. 6). Rain convected heat was minimal, mainly because of the low rainfall magnitude and partly because the rain temperature (approximated by the air temperature) was often not greatly different from the near surface soil temperature. Over the study period, the incoming ground heat flux to the fen soil totalled 106 MJ, compared with 100 MJ that entered the polar desert soil (Table 3).

Out-going heat flux from the active layer to the permafrost below, computed using Eq. (3), was large for both the fen and the polar desert soils (Fig. 6). Daily fluctuations were not large and the seasonal totals were 44 MJ for the fen and 46 MJ for the polar desert, representing 41 and 46 per cents of their respective incoming ground heat fluxes (Table 3). Warming of the active layer at both sites (calculated

Table 3 – Active layer heat balance, July 2-August 16, 1988

	Polar desert site		Fen site	
	MJ	%	MJ	%
Q_g				
conducted	99.5	100	106.4	100
convected	-0.04		-0.1	
Q_l	48.7	49	60.2	57
Q_s	5.0	5	2.4	2
Q_p	45.8	46	43.7	41

Q_g – incoming ground heat
Q_l – latent heat for ground thaw
Q_s – sensible heat for active layer warming and cooling
Q_p – heat conducted to permafrost

using Eq. (4)) accounted for small quantities of heat. Over the thawed period, 2 MJ (or 2 per cent of incoming heat flux) was used for active layer warming at the fen and it was 5 MJ (or 5 per cent of incoming heat) at the polar desert site.

The latent heat used for ground thaw was computed as the residual term of Eq. (1) and its cumulative values are

$$\int_0^\tau Q_1 = \int_0^\tau Q_g dt - \int_0^\tau Q_s dt - \int_0^\tau Q_p dt \tag{8}$$

where 0 is the beginning of the thaw season (2 July) and =1,46 days. Various components of Eq. (9) are plotted as Fig. 7.

Latent heat was large at the beginning of the thaw season, corresponding with the period of rapid thawing of the active layer. Later, latent heat consumption was reduced as ground thaw slackened. Over the study period, Q_l calculated as a residual of the energy balance (Eq. (8)) was 60 MJ for the fen site, and 49 MJ for the polar desert, representing 57 and 49 per cent of their respective incoming ground heat flux (Table 3). These values compared favourably with the Q_i calculated using frost table depth and ice content of the soils (Eq. (5)) which yielded 51.5 ± 7.7 MJ for the fen and 41.6 ± 3.2 MJ for the polar desert. Such results demonstrate that latent heat consumption is a significant component of the active layer heat balance and this is a notable feature of permafrost soils (Roulet and Woo 1986; Rouse 1982).

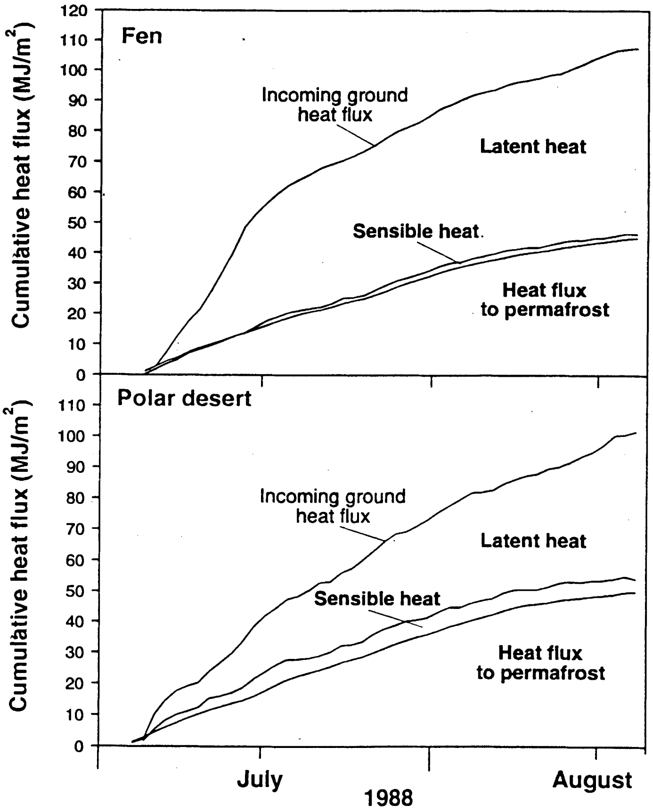

Fig. 7. Cumulative heat balance of the active layer during the thaw season of 1988, fen and polar desert soils.

Discussion and Conclusions

1) Conduction has been found to be the main mechanism of heat transfer in frozen soil (Nixon 1975). Thermal conductivity of the soil material therefore is an important factor, but it varies with the moisture content, both in frozen and unfrozen states. Hinzman *et al.* (1991) demonstrated this by laboratory measurements of the effective thermal conductivity as a function of soil type, temperature and moisture content. The thermal conductivity of our soils, computed using empirical equations, compared well with those provided by Hinzman *et al.* (their Fig. 12). This lends confidence to our calculated values. Our study shows that for saturated soils, thawing causes sharp switch in the thermal properties (including conductivity and heat capacity, hence also the thermal diffusivity which is K/C). For soils which experience moisture fluctuations, the thermal parameters are also sensitive to drying and wetting events. Thus, the moisture dynamics of the active layer strongly influence a heat conduction through the effects of the thermal properties.

2) In terms of active layer thaw, various studies have linked the thawing depth to the soil thermal properties and the energy flux into the soil. Rouse *et al.* (1992) measured deeper thawing depth in a dry season than a wet one because of larger ground heat flux and higher thermal diffusivity. Nakano and Brown (1992) found an increase in thaw depth as the thermal diffusivity increases. Similarly, Smith (1977) observed that a site with larger thermal diffusivity led to deeper thawing than its adjacent wet site. None of these studies related the thawing depth explicitly to the ground ice content. However, active layer thaw necessarily involves both the warming of the soil materials to freezing temperatures and a phase change for the ice in the soil. Using field data, our study demonstrates the considerable amount of latent heat consumed by the active layer being an order of magnitude larger than the heat that warms the soil. The role of ground ice in affecting active layer thaw was revealed by contrasting the heat fluxes in two soils with diferent ice contents. For the fen soil with large ice content throughout its profile, its maximum seasonal thaw depth was only 60% of the less ice polar desert soil. Evidently, ground ice content strongly controls active layer thawing depth.

It is further surmised that soils in poorly drained sites are often rich in ice to retard the development of a deep active layer. Shallow thaw confines the summer moisture within a shallow zone to augment the ice content when the ground freezes. Such a feature is common in the Arctic wetlands and indicates a feedback between the hydrological and the thermal aspects of the active layer.

Acknowledgements

This work was funded by the Natural Sciences and Engineering Research Council of Canada and the Polar Continental Shelf Project provided generous logistical support. We thank Dr. Philip Marsh for his help and we acknowledge the field assistance of Mary Ferguson and Philip Giles.

References

Cruickshank, J. (1971) Soil and terrain units around Resolute, Cornwallis Island, *Arctic, Vol. 24*, pp. 195-209.

de Vries, D. A. (1963) Thermal properties of soils. In : W. R. van Wijk (Ed.), *Physics of Plant Environment*, North-Holland Publishing Co., Amsterdam, pp. 210-235

Farouki, O. T. (1981) Thermal properties of soils, CRREL Monograph No. 81-1, 136 pp.

Hinzman, L. D., Kane, D. L., Gieck, R. E., and Everett, K. R. (1991) Hydrologic and thermal properties of the active layer in the Alaskan Arctic, *Cold Regions Science and Technology, Vol. 19*, pp. 95-110.

Nakano, Y., and Brown, J. (1972). Mathematical modeling and validation of the thermal regimes in tundra soils, Barrow, Alaska, *Arctic and Alpine Research, Vol. 4*, pp. 19-38.

Nixon, J. F. (1975) The role of convective heat transport in the thawing of frozen soil , *Canadian Geotechnic Journal, Vol. 12*, pp. 425-429.

Roulet, N. T., and Woo, M. K. (1986) Wetland and lake evaporation in the low Arctic, *Arctic and Alpine Research, Vol. 18*, pp. 195-200.

Rouse, W. R. (1982) Microclimate of low Arctic tundra and forest at Churchill, Manitoba. In: H. M. French (Ed.), Proceedings of Fourth Canadian Permafrost Conference, National Research Council of Canada, Ottawa, pp. 68-80

Rouse, W. R., Carlson, D.W., and Werck, E. J. (1992) Impacts of summer warming on the energy and water balance of wetland tundra, *Climate Change, Vol. 22*, pp. 305-326.

Smith, M. W. (1977) Computer simulations of microclimate and ground thermal regimes: test results and program description, ALUR Report 75-76-72, Department of Indian and Northern Affairs, Ottawa, 74 pp.

Woo, M. K. (1986) Permafrost hydrology in North America, *Atmosphere-Ocean, Vol. 24*, pp. 201-234.

Woo, M. K., and Marsh, P. (1990) Response of soil moisture change to hydrological processes in a continuous permafrost environment, *Nordic Hydrology, Vol. 21*, pp. 235-252.

First received: April, 1995
Revised version received: October, 1995
Accepted: 27 November, 1995

Address:
Geography Department,
McMaster University,
Hamilton, Ont. L8S 4K1,
Canada.