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Response of Soil Moisture Change to Hydrological Processes in a Continuous Permafrost Environment

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The moisture content of the active layer at three sites in a continuous permafrost area was measured using a twin-probe gamma density meter. The moisture storage status at these sites were related to various hydrological processes. Moisture was gained by meltwater and rainfall infiltration, but lost to evaporation in summer. Lateral inflow maintained a thick saturated zone at the fen (wetland) site. At the gravel site, there was a net moisture loss due to evaporation and lateral outflow.

Moisture changes in the active layer during the summer were examined in terms of the water balance at the three sites. This established quantitative relationships between the moisture regime and the major hydrological processes in the permafrost environment.

Introduction

Ecological studies of the permafrost environment, and modelling of permafrost hydrological regimes require good appreciation of the moisture distribution in the active layer. Fig. 1 provides a conceptual framework, integrating the various processes that contribute water or extract moisture from the Arctic soils, together with the mechanisms that redistribute moisture in the soil profile. Meltwater and rainfall are the major sources of water inputs, though on sloping sites, lateral flow is also important. Evaporation and, in some cases, lateral flow are the main mechanisms of water loss. In response to these external water inputs and losses, and to the thermal gradient at the freezing front, moisture moves up or down the soil column.

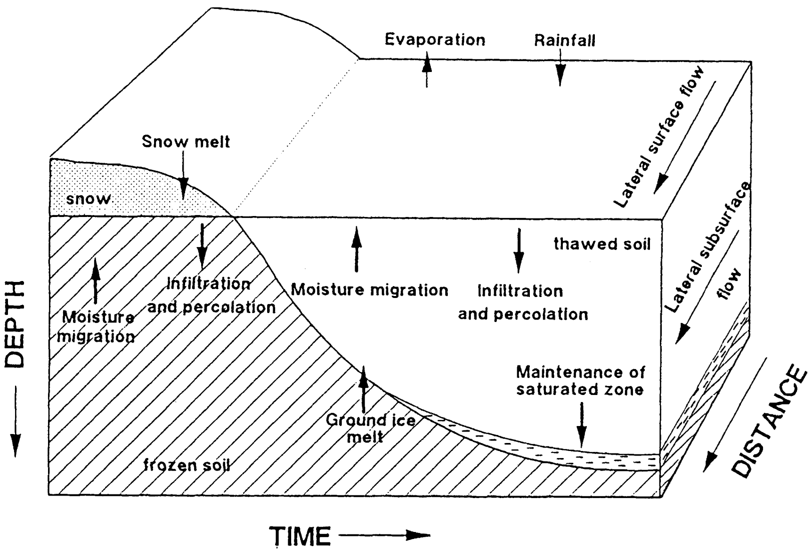

Fig. 1. Conceptualized diagram showing the external inputs and losses (thin arrows) and internal processes (thick arrows) that affect the soil moisture change in the active layer.

This can occur in the frozen and in the thawed soils.

In the past, hydrological studies have focused upon single aspects of soil moisture or groundwater movement in the active layer. Examples include Guymon and Luthin (1974), Kane and Stein (1983) Marsh (1988), Woo and Steer (1983). Only Guymon (1975, 1976) has produced integrated surveys of moisture change in Arctic soils as affected by various thermal and hydrological processes in the course of a year. This paper identifies and discusses the various external forces and internal mechanisms that govern the moisture distribution in vertical soil columns. Three types of soil materials commonly encountered in continuous permafrost areas are selected for comparison. The results will indicate the relative magnitudes of various processes influencing the moisture regimes of the active layer.

Study Site

Field work was carried out in 1988-89 near Resolute, Northwest Territories (74° 55', 94° 51'W) at three sites which are within 1 km of each other. These sites represent three typical surficial materials found in the area (Cruickshank 1971). They include polar desert soil which is a mixture of pebbles and loam (site A), bog or wetland soil (site B), and lithosol or gravelly materials (site C). The »bog« soil occupies a fen that is fed by snowmelt throughout the entire summer. In this study it is referred to as the fen site.

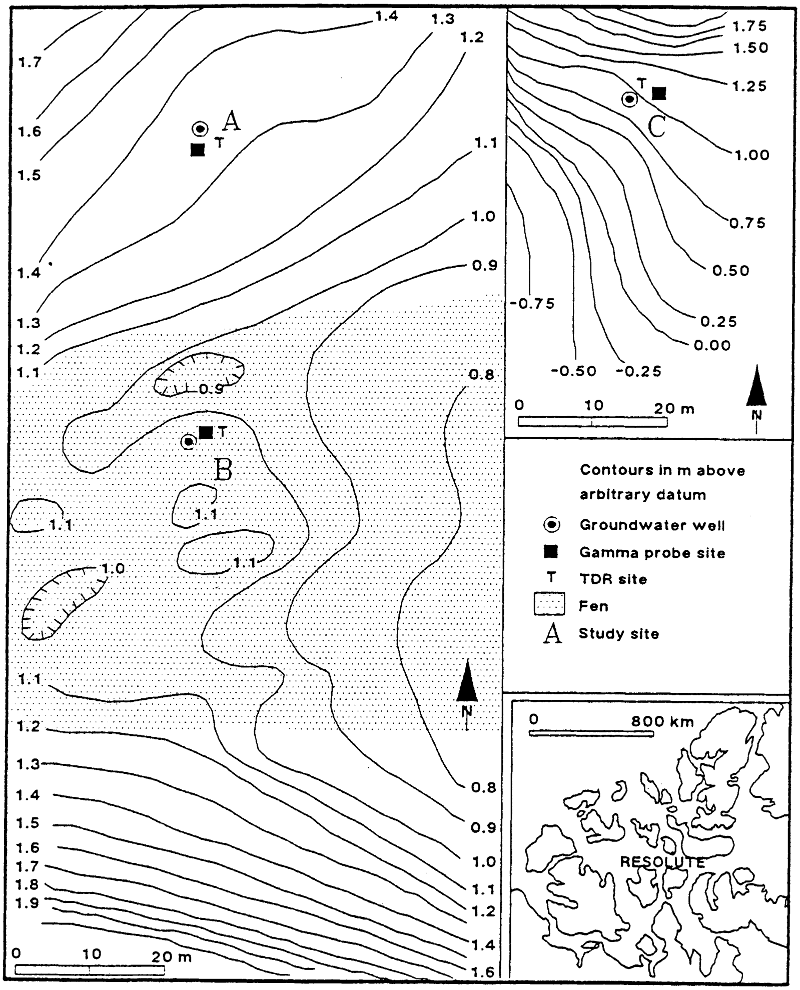

Fig. 2. Topography and instrumentation at the three study sites: A) Polar desert
B) Fen
C) Gravel

As shown by the contours (Fig. 2), the polar desert site has a surface slope of 0.01 and its surface is bare except for a few scattered clumps of purple saxifrage. Normally, the permafrost table occurs at a depth of 0.45 m, but during the exceptionally warm summer of 1988, permafrost receded to 0.55 m. The fen site is fed by meltwater from late-lying snowbanks occupying the slope west of the site (the steeper but dry slope to its south is not a notable source of water input). Despite the low west to east gradient of 0.01, there is noticeable surface and subsurface flows through the fen throughout summer. Its surface is completely vegetated by

Table 1 – Grain size distribution and hydraulic characteristics of three Resolute soils

	Clastic Portion				Bulk	Organic	Hydraulic
	Clay %	Silt %	Sand %	Gravel %	density ($kg.m^{-3}$)	content (%)	conductivity ($m.s^{-1}$)
Polar desert:	9	19	42	30	1709	2	2.4×10^{-4}
Fen:	21	53	23	3	1683	15	–
Gravel:	3	9	18	70	1709	0	1.9×10^{-1}

grasses, sedges, and mosses. A thin peat layer (0.05 m) overlies loamy soil at this site and the permafrost occurs at a depth of 0.37 m. The gravel site which occupies a hollow on a relatively steep but short slope (surface gradient is 0.06), is frequently covered by a deep snowdrift in spring. The surface is completely barren, with the permafrost table normally at 0.65 m depth, but it reached 0.7 m by the end of the unusually warm 1988 summer. The grain size distribution of these three soils, together with their hydraulic characteristics, is summarized in Table 1.

A comparison of the air temperature during the field season with the 30-year mean from Resolute weather station shows that much milder conditions prevailed in 1988 (Fig. 3). Snow accumulation varied among the three sites, with depths of 0.43, 0.47, and 0.64 m at the polar desert, fen and gravel sites respectively. Snowmelt increased gradually, as the net radiation increased (due to the decreasing albedo), and as the air temperature rose. By July 1, the presence of snow-free patches led to heat advection from the bare ground to the snow, and high melt rates were observed at the residual snow cover. Between June 30 and August 23, rainfall totalled 21 mm and this amount is within the usual range of summer rainfall.

Instrumentation

At the three experimental sites, gamma access tubes were installed down to the permafrost table in late August 1988. Changes in the frozen and unfrozen water stored in the active layer were determined using a twin-probe gamma density meter. Measurements began in June 1988 and continued at frequent intervals throughout the summer. Moisture profiles were obtained at 5 cm increments, with each measurement accurate to ± 2 mm of water (Marsh 1988). At each site, snow depth was monitored during the melt period; and afterwards water table position was measured from a well excavated in the active layer. The frost table position was monitored daily or once every two days, and ground temperatures at each site were measured using thermistors inserted at 2, 5, 10, 25 and 50 cm depths.

A meteorological station was set up midway between the polar desert and the gravel sites to measure air temperature, relative humidity, wind speed, solar radiation and rainfall. These data, recorded at hourly intervals by a Campbell CR 21 data logger, are considered to be representative of the three study sites because of their proximity. Net radiation was measured over snow, fen and polar desert sites. The meteorological data thus obtained were used to compute snowmelt and evaporation, and the methods are given in the relevant sections to follow.

Active Layer Moisture Change

For this study, all moisture changes determined from the gamma measurements are compared to the initial moisture conditions for June 20, 1988. Thus, at a time T days after June 20, the moisture change at a particular depth z is $ds(z,T)$. For the vertical column comprising the entire depth Z of the gamma tubes the soil moisture change is

$$\Delta S'(T) = \int_0^Z ds(z)\,dz\Big|_T \tag{1}$$

where $\Delta S'(T)$ is storage change for the entire column between day 0 and day T.

Although the probes at the gravel site did not fully penetrate the thawed zone for 1988 because of exceptionally deep thawing, the base of the thawed zone was often saturated. The thawed zone can be divided into a saturated and a non-saturated section, separated by the water table whose position was measured every other day. The non-saturated section was shallow enough to be measured by the gamma probe and the saturated moisture change, $\Delta s(sat)$, was estimated to be 7 mm per unit depth for the gravel site (based on the measurements made on July 10 when the saturated zone was within the depth reached by the gamma probe). Then, moisture change between day 0 and day T for the thawed gravel column is

$$\Delta S(T) \equiv \int_0^W ds(z)\,dz\Big|_T + \int_W^F \Delta s(sat)\,dz\Big|_T \tag{2}$$

where W and F are the positions of the water table and the frost table, and $\Delta S(T)$ is storage change for the maximum thawed zone.

Using Eq. (1) for the polar desert and fen sites, and Eq. (2) for the gravel site, the soil moisture change for various days during the study season was obtained (Fig. 3). At all sites, there was a moisture gain during the snowmelt period, followed by a decline during the dry post-melt days. The polar desert had a positive change in active layer moisture after the August rain events, while the gravel site had a decline from early July and then steadied off after mid-July. The fen site showed a gradual moisture gain during the latter half of summer. In the next section, these patterns of moisture change will be examined in conjunction with the external hydrological inputs and losses imposed upon the active layer column of the three sites.

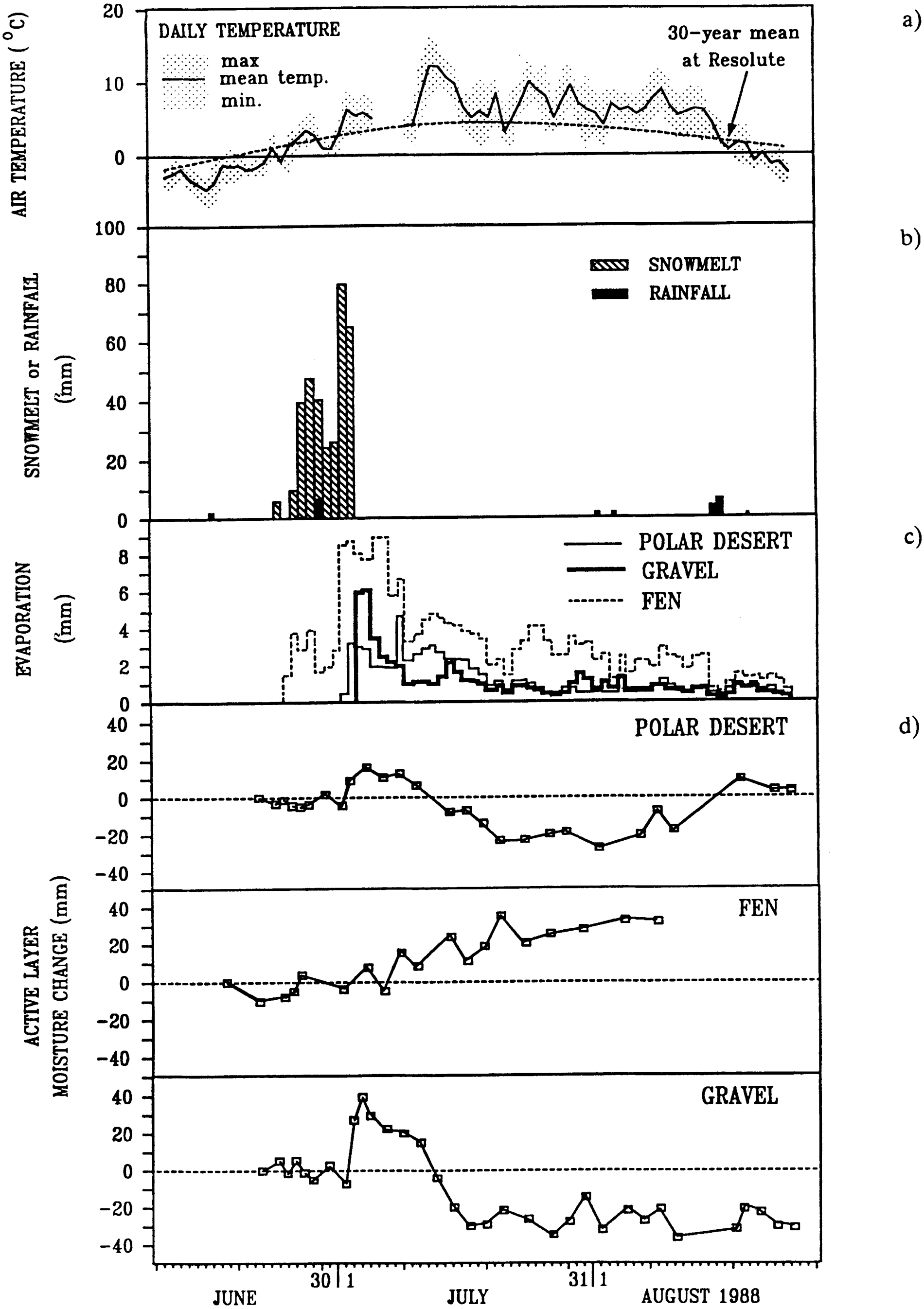

Fig. 3. a) Air temperature, b) snowmelt and rainfall, c) evaporation and d) active layer moisture change at the three study sites during the 1988 field season.

Active Layer Water Gains and Losses

Snowmelt and Rainfall

Unusually early snowmelt in May 1988 was followed by heavy snowfall and low temperatures which effectively refroze the meltwater produced by these melt events. The main melt season arrived in late June, and proceeded rapidly. Fig. 3 shows the daily snowmelt computed by the energy balance approach (Heron and Woo 1978) and verified by snow ablation measurements in the field. The snowmelt computation methods are identical to those described by Moore (1983) and Price and Dunne (1976) and will not be repeated here.

Abundant quantities of meltwater were released between June 26-30, and the only limiting factor for soil moisture replenishment was the low infiltration capacity of the frozen soil (*cf.* Kane and Stein 1983). Any water that was unable to infiltrate was refrozen at the base of the snow to form basal ice (March and Woo 1984), or ran off as surface flow from the sites. The rise of the »water table« to maintain a free water surface above the ground between June 26-30 (Fig. 4) bore evidence to this effect.

Rainfall in the High Arctic is seldom of high intensity. The amount of rain that fell in 1988 (21 mm) was substantially less than snow accumulation at the study sites. However, rainfall events were often able to increase the soil moisture storage. This was particularly noticeable for the 11 mm rain which fell between August 17 and 18.

Evaporation

The Priestley and Taylor (1972) model has been found to be suitable for calculating evaporation for Arctic sites (Marsh *et al.* 1981)

$$E = \frac{\alpha\, \sigma\, (Q^{*} - Q_g)}{(\sigma + \gamma)\, \rho\, \lambda} \qquad (3)$$

where E is evaporation in m/s, σ is the slope of saturation vapour pressure *vs* temperature curve, Q^* is net radiation, Q_g is ground heat flux, both in W/m^2, γ is the psychrometric constant, ϱ is density of water (kg/m^3) and λ is the latent heat of fusion (J/kg). For saturated conditions, $\alpha = 1.26$ is appropriate, and this has been applied to the estimation of evaporation from wetlands (Rouse *et al.* 1977). For the polar dessert and gravel soils, α is related to soil moisture content (Marsh *et al.* 1981)

$$\alpha = \frac{1.26}{\exp(5.24 - 21.56M) + 1} \qquad (4)$$

where M is the surface (0 to 6 cm) volumetric moisture content in decimal fractions.

Evaporation at the three sites were calculated using Eq. (3), and their daily values are presented in Fig. 3. At the fen site, an $\alpha = 1.26$ was used as the water

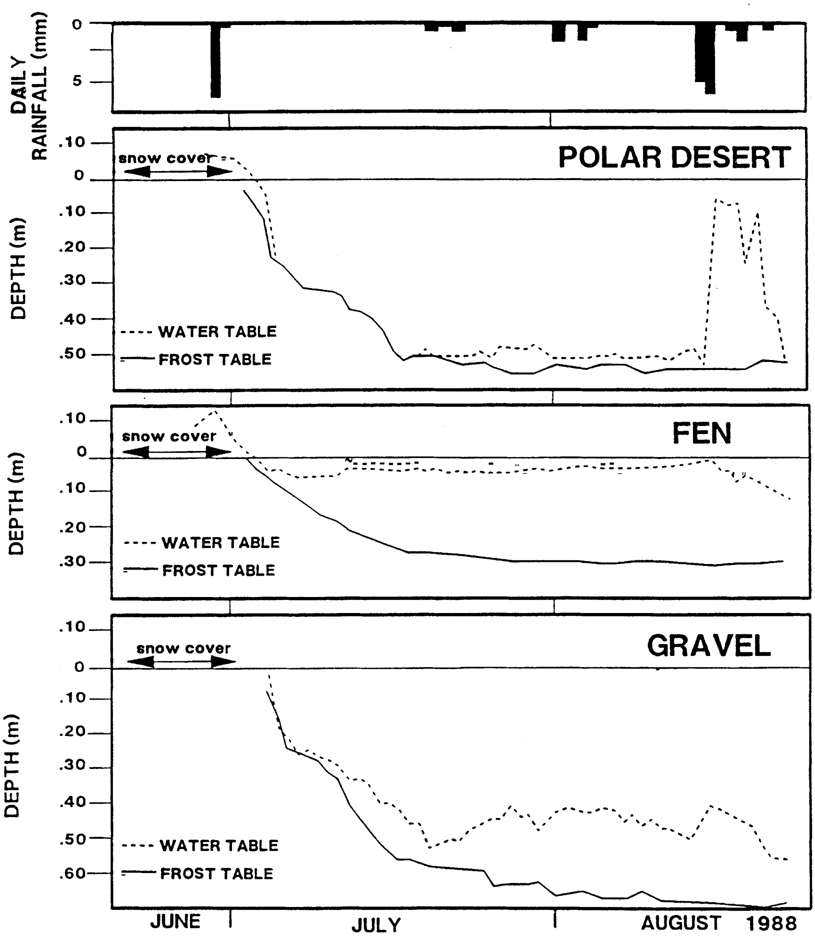

Fig. 4. Changes in water table and frost table positions during the summer of 1988 at three study sites.

table was always close to the surface. The computed evaporation was the highest at this site. The polar desert, with more moisture at the surface than the gravel, had the next highest amount of evaporation. At all the sites, evaporation accounted for a significant portion of water loss during the thawed season (Table 2).

Lateral Flow

The observed occurrence of surface runoff at the fen site and the high gradient at the gravel site suggest that lateral flow was important. The net gain or loss of water at a site due to lateral flow can be obtained from the water balance relationship

$$\int_0^T Q\,dt = \Delta S(T) - \int_0^T (R-E)\,dt \qquad (5)$$

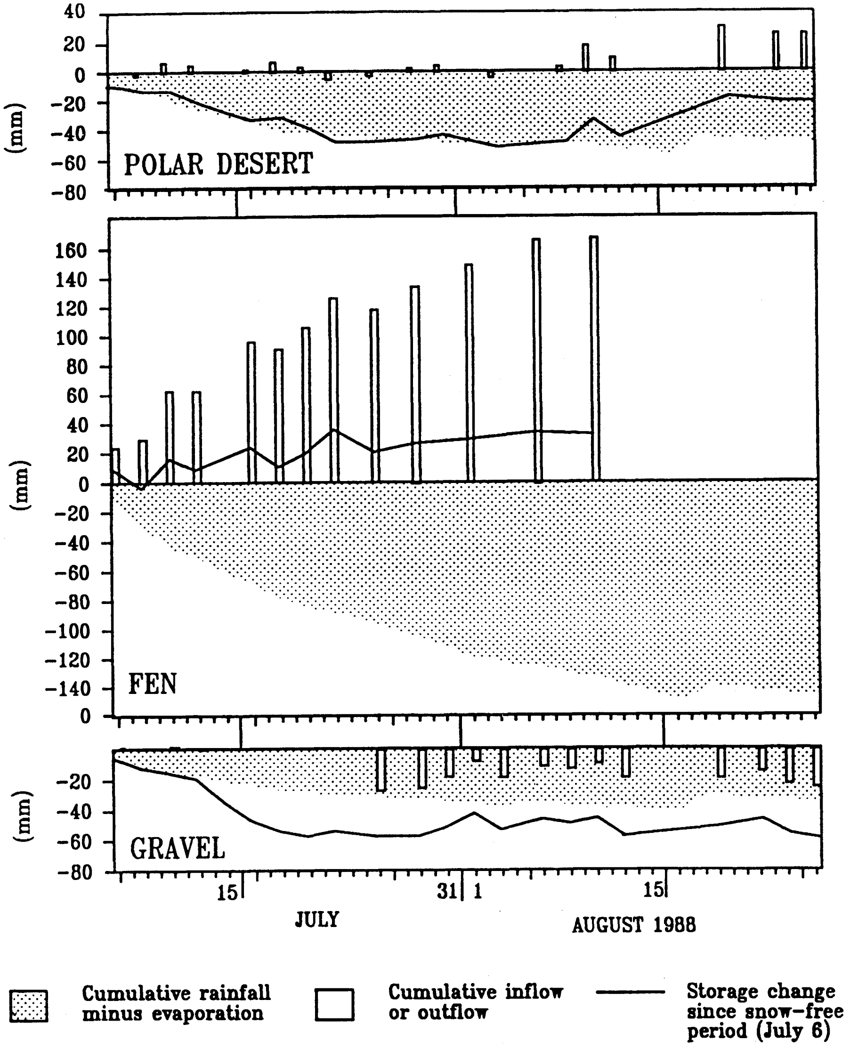

Fig. 5. Water balance and computed cumulative inflow or outflow at the three sites (see text for explanation).

where Q is lateral flow to or from a soil column, R is rainfall, E is evaporation obtained from Eq. (3) and $\Delta S(T)$ is moisture storage change between day 0 and day T.

Fig. 5 showed the computed cumulative lateral flows to and from the three sites. Flow contribution was largest to the fen site, especially in early melt season. The gravel site had a net loss of lateral flow to supply water to the slope below it. Net lateral outflow and evaporation caused a moisture storage deficit at this site.

Moisture Distribution within the Active Layer

Moisture in the active layer is subject to redistribution. Figs. 6, 7, and 8 shows the changes in soil moisture at various depths during different times in the study period. These diagrams show the values of $ds(z,t)$ for various depths and on selected days t. The measured points are shown as symbols and a ± 2 mm error band shows the uncertainties of the gamma measurements.

Infiltration into Frozen Soil

Kane and Stein (1983) and Gray *et al.* (1986) have measured the infiltration of meltwater into frozen soils. At the Resolute sites, such a phenomenon can be seen from the moisture change within the vertical profiles.

When snowmelt began on June 23, 1988, the soil was at about −8° C. There was indirect evidence of infiltration as the soil temperatures rose sharply by about 3° within one day, at all depths down to 0.3-0.5 m (Fig. 9). This suggests a large latent heat transfer to the frozen soil as the infiltrated meltwater refroze (Woo and Heron 1981). Infiltration was especially pronounced at the gravel site where the July 3 moisture change profile showed 6 to 8 mm increases at around 0.25 m depth (Fig. 8). The gravel has a large hydraulic conductivity when frozen because its pore space was probably not entirely blocked by ice; and infiltration was less impeded than at the polar desert site. The fen site showed only a minor moisture increase in its profile, and this may be due to the highly saturated nature of the soil which, when frozen, yielded ample ice to seal the pores.

Moisture Depletion in Thawed Soil

Vertical drainage and evaporation depleted the moisture in the upper thawed zone of the active layer. The former process was evident when comparing the moisture profiles at the gravel site between July 10 and 26, as thawing deepened and moisture drained from the upper zone (Fig. 8). Vertical drainage was less efficient at the polar desert site because of lower permeability (Table 1). For example, between July 13 and 20, soil moisture at 0.05-0.20 m dropped by only 2 mm (Fig. 6).

Near the surface, moisture loss was noticeable at all three sites during the dry period in mid-July. Evaporation was the prime cause of water loss at the top 0.02 m, though the exact vertical extent of the active layer which was affected by evaporative loss is unknown.

Infiltration into Thawed Soil

Rainfall and surface inflow are the major sources of water that infiltrate the thawed soil. Rainfall events in mid-August 1988 raised the moisture content at the surface zone of all sites (*eg.* the August 8 profiles for polar desert site). At the fen site, lateral surface inflow from the west provided a steady supply of water. This water infiltrated the porous peat layer easily. During rainfall and surface inflow events,

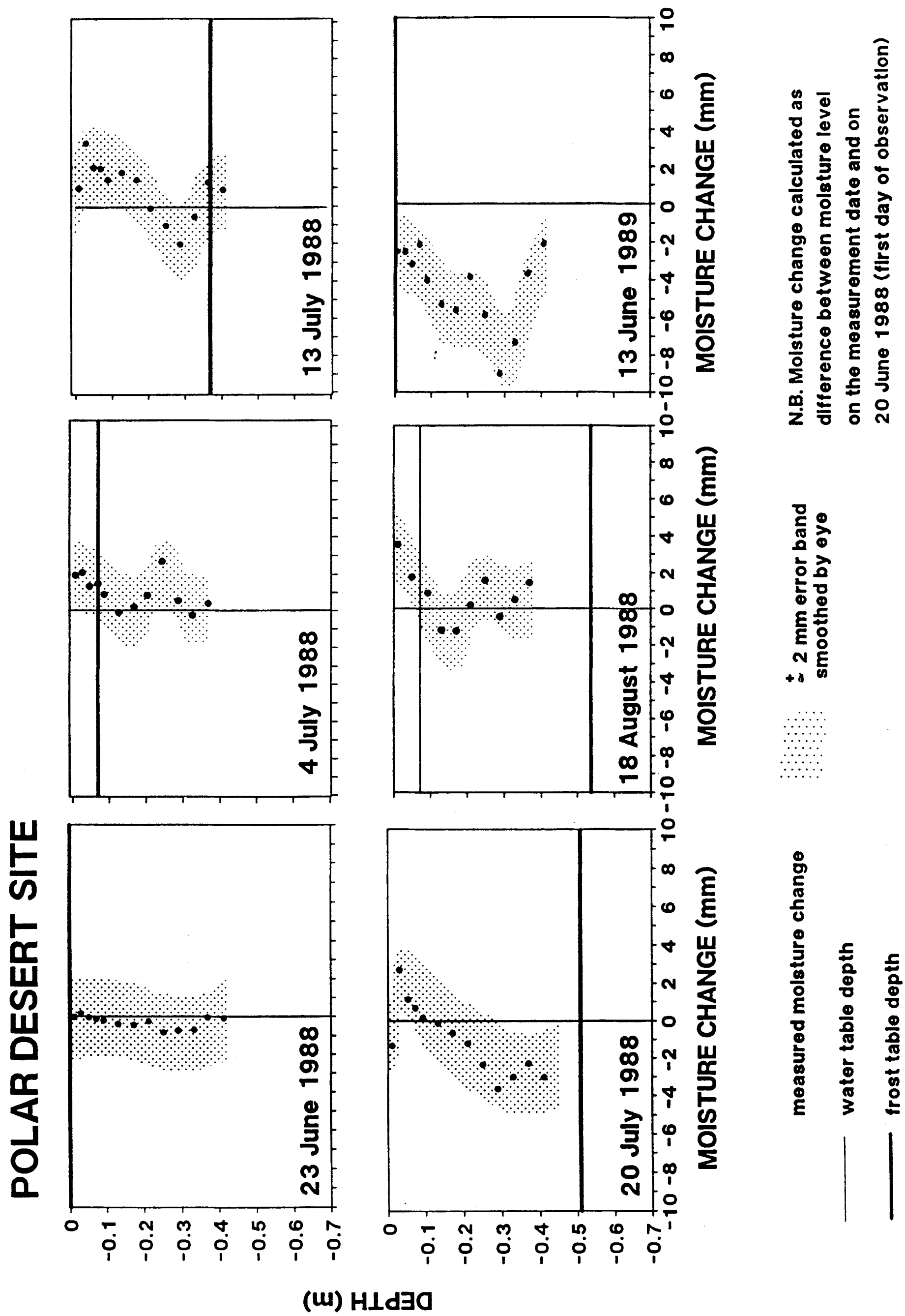

Fig. 6. Soil moisture changes at the polar desert site on selected days in 1988 and 1989.

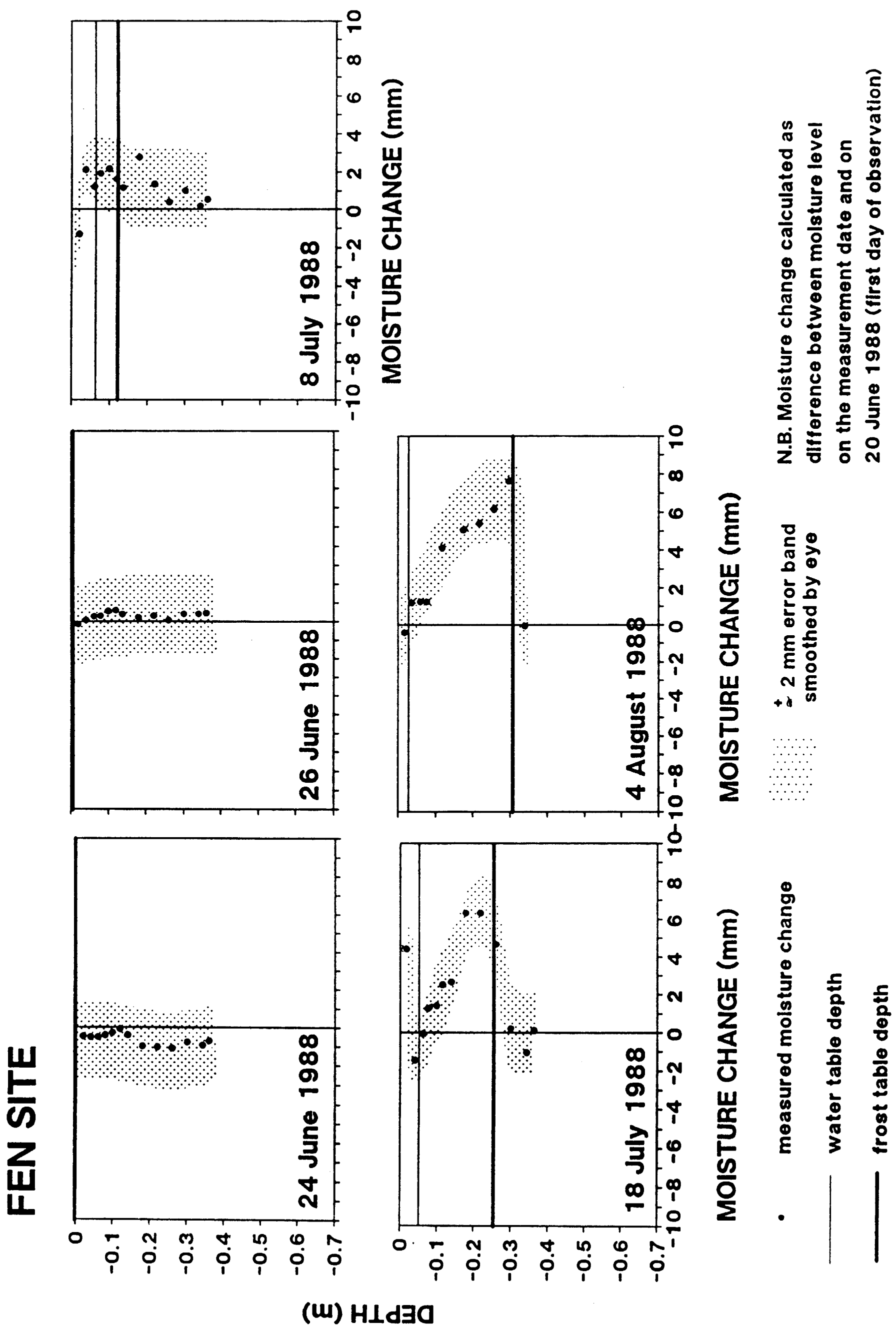

Fig. 7. Soil moisture changes at the fen site on selected days in 1988 and 1989.

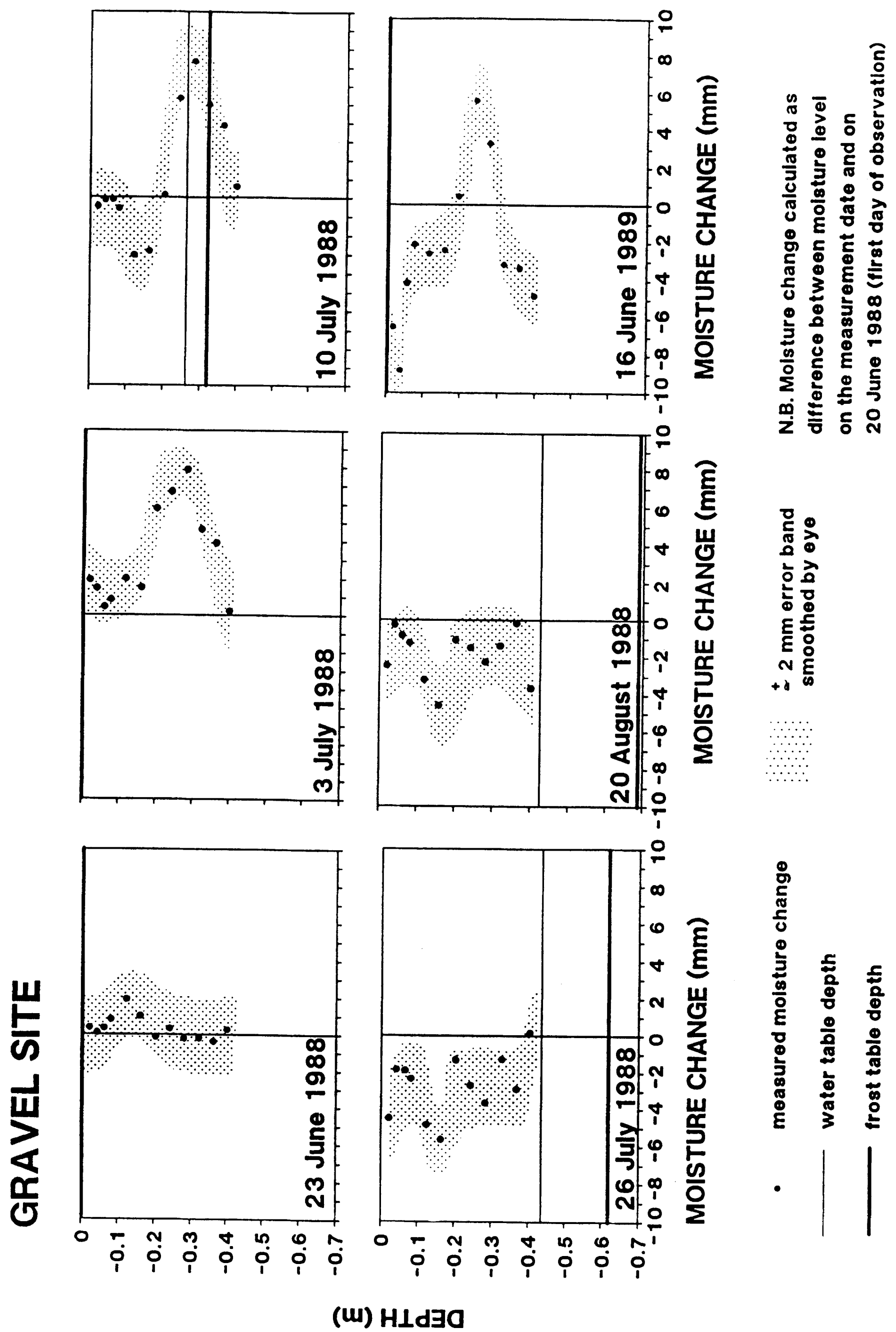

Fig. 8. Soil moisture changes at the gravel site on selected days in 1988 and 1989.

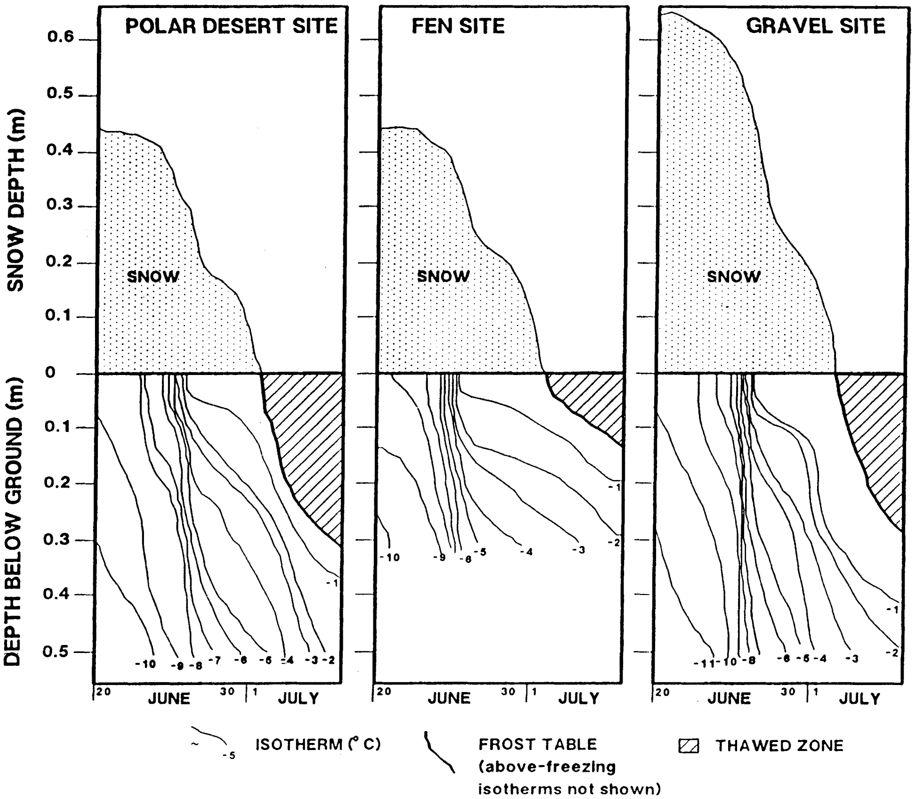

Fig. 9. Snow depths and ground temperatures at the three study sites during the snowmelt period of 1988. Rapid rises in ground temperatures suggest infiltration followed by refreezing of meltwater in frozen soils.

moisture content at the near-surface zone rose quickly but afterwards, good drainage and evaporation reduced the moisture equally fast. The surface layer therefore experienced moisture fluctuations during the thaw period (Fig. 7).

Fluctuation of the Saturated Zone

The maintenance of a saturated (suprapermafrost groundwater) zone requires adequate moisture supply and a relatively impermeable substrate to prevent downward loss through drainage. It is clear from the July 10 profile in Fig. 8 that even gravels have very low permeability below the frost table in summer, probably because most of the pores are sealed by ice to retard percolation. As the frost table descended, the ground ice would melt and become a source of water to the saturated zone. This constitutes only a change of state, and does not affect the soil moisture change.

Unless fed by additional water supply, the saturated zone moves deeper in the active layer as the frost table descends. Fig. 4 shows the position of the water table at the three sites. The water table decline in early July illustrated the effect of a falling frost table. Water table rises were often associated with the percolation of rainwater.

During the melt season, water table at the polar desert and fen sites rose to flood the ground because meltwater supply exceeded the infiltration capacity of the frozen soil. At the fen site, a high water table was sustained in summer because of lateral inflow, as explained in the last section. It is this water source that created the wetland. Indeed, the wetland shed water to both its northern and southern fringes because its water table was higher than the non-wetland polar deserts immediately adjacent to it.

Lateral flow was much less important at the polar desert site, and a saturated zone did not occur during the rain-free period of early and mid-July (Fig. 4). Later, a thin saturated zone developed, which expanded rapidly during the rain event of August 16-17. The magnitude of this rise was too high to be sustained by rainfall alone, and some lateral inflow was probably responsible. At the gravel site, a saturated zone persisted throughout the study period. This zone thickened after mid-July and was maintained for the rest of the summer by both rainfall and by lateral flow.

Moisture Migration in Frozen Soil

The migration of moisture to the freezing front is a well documented phenomenon (Anderson *et al.* 1984). One consequence is the upward migration of moisture in early winter, so that the near-surface zone is enriched while the middle portion is desiccated correspondingly (Guymon 1976). Santeford (1978), Woo (1982), Smith and Burn (1987) also demonstrated that there can be moisture loss from the soil through vapour flux to the overlying snow. Upward moisture migration to the freezing front and desiccation of the zone below were noticeably demonstrated by the June 16, 1989 moisture profile of the frost susceptible, polar desert soil.

Discussion and Conclusion

The water balance framework allows the partitioning of the active layer moisture storage, $\Delta S(T)$, into its contributing components. For the post-melt period

$$\Delta S(T) \equiv \int_{t_0}^{T} (R-E+Q)\,dt \tag{6}$$

where R is rainfall, E is evaporation, Q is lateral flow which may be a net gain or a net loss. The relative importance of the various components at the three sites are compared for the period t_o = June 20, 1988, and T = July 8 or July 10. The

Table 2 = Water balance at three study sties during the post snowmelt dry period of 1988. All values given in mm

	Polar desert	Fen	Gravel
Water balance period	July 5-Aug 8	July 5-Aug 10	July 5-Aug 8
Storage change	-46	32	-49
Rainfall	7	7	7
Evaporation	56	142	43
Lateral flow	negligible	167*	-13*

Note: * values are calculated as residuals of the water balance and no error term is included in the calculation: storage change = rainfall - evaporation + lateral flow

beginning time corresponds to when the snow disappeared and the ending time was when the gamma access tube at the fen site became flooded and no more measurements were possible.

Table 2 shows that the active layer storage experienced a net loss for both the polar desert and the gravel sites, but a net gain for the fen site. The loss at the polar desert site was largely due to evaporation, and for the gravel site, it was caused by both evaporation and net loss to lateral flow. The computed evaporation loss was large for the fen site, but this was more than made up by the net gain in lateral inflow. During this period, rainfall was low. The values given in Table 2 did not include error estimates, though even if a 10 per cent error is assumed, the relative contribution of the various components remains unchanged.

These results confirm quantitatively the relative importance of various processes that influence the moisture regime of the active layer, as depicted in Fig. 1. There are also notable difference amongst the three types of sites. During the melt period, infiltration and percolation in the frozen soil may enrich part of the vertical profile, and the rate was related to the perviousness of the frozen soil. Gravels provided the largest permeability and the frozen wetland soil was the least permeable because of the abundant ice in its pores. During the relatively dry post-melt period, the polar desert site was essentially exchanging moisture only with the atmosphere, through rainfall and evaporation. The gravel site, located on a slope, gradually yielded lateral flow from its moisture storage to result in a net storage loss at the end of the summer. The fen site gained considerable lateral flow from meltwater released by late-lying snowbanks, and this was able to sustain high evaporation and maintain positive moisture storage, thus giving rise to a wetland environment.

In summary, this paper has demonstrated that changes in the soil moisture content of the active layer are affected by 1) the intrinsic water holding and transmission properties of the materials, both under a frozen and a thawed state 2) the seasonal variations of the hydrological activities external to the soil column, such as snowmelt, rainfall, evaporation and lateral flow 3) the thermal and hydrological processes operating within the active layer, leading to redistribution of moisture along the vertical profile.

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