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Snowmelt Hydrology of Two Subarctic Slopes, Southern Yukon, Canada

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Large quantities of water are discharged from subarctic basins during snowmelt season. Runoff contributing areas as well as timing and magnitude of meltwater generation from different slopes are highly variable. Two slopes in the lower Wolf Creek basin, southern Yukon, were studied in 1997. The south-facing slope has a dense aspen forest that is leafless in the melt period (April – May) and is underlain by seasonal frost. The north-facing slope has open stands of spruce and an organic layer that rests on mineral soils with permafrost. In 1997, snowmelt is advanced by over 10 days on the south slope, which receives more solar radiation than the north aspect. All meltwater on the south slope infiltrates the frozen silt without generating runoff. By the time significant melt events occur on the north slope, the frost and snow are gone from the south. Meltwater is able to infiltrate the frozen organic soil but deep percolation is prevented by the ice-rich substrate. Lateral flow begins after the organic layer is saturated, with much runoff along intermittent rills fed by diffuse and pipe flows. Rills and pipes are interconnected but the drainage network and runoff contributing area change depending on the disposition of the snow as well as water and frost table positions relative to local topography. Contrasts between the north and south slopes have important implications on direct runoff generation during the melt period. Situations similar to the study site can be found elsewhere in subarctic North America and the observed processes have a bearing upon hydrological modelling for the subarctic environment.

Introduction

In the subarctic environment, snowmelt yields large amounts of water to streamflow. Heterogeneity within subarctic basins produces considerable variations in the flow characteristics of headwater catchments. For instance, Slaughter *et al.* (1983) noted that flashier flows are associated with catchments with large percentages of permafrost. Research in central Alaska (Dingman 1973; Kane *et al.* 1981) reported the occurrence of snowmelt runoff from organic-covered, permafrost slopes. On the other hand, Kane *et al.* (1981) and Gibson *et al.* (1993) observed that non-permafrost slopes contributed little to streamflow. An examination of subarctic hillslopes with different orientations indicates physical contrasts which should have a bearing upon their hydrological characteristics. In particular, slopes of north and south aspects are highly different in: 1) radiation regime, 2) vegetation (spruce *vs.* birch-aspen forests), 3) soil surface cover (organic mat *vs.* leaf litter) and 4) frozen ground conditions (permafrost *vs.* seasonal frost). These features affect the snowmelt and meltwater delivery processes on the slopes.

Investigations of hydrological processes have been carried out in subarctic sites underlain by seasonal frost or permafrost, ranging in scale from experimental plots to entire drainage basins (Chacho and Bredthauer 1982; Dingman 1966, 1973; Kane *et al.* 1981; Kane and Stein 1983; Santeford 1979a; Wright 1979). Energy balance approach is frequently used to study snowmelt processes (Eaton and Wendler 1982; Price and Dunne 1976). Infiltration into frozen soils, be they seasonal frost or permafrost, has been measured in Central Alaska (Kane *et al.* 1981; Kane and Stein 1983). The hydrological behaviour of the organic layer on permafrost slopes has often been emphasized (Slaughter and Kane 1979) particularly because most snowmelt runoff is transmitted down the slopes in these soils (*cf.* Hinzman *et al.* 1993).

Results from these studies have improved our understanding of subarctic hydrology. On the other hand, there are no direct inter-slope comparisons regarding their hydrological responses to the snowmelt and runoff processes. Since drainage basins comprise slopes of different orientations, contrasts in quantity, timing and mechanisms of water delivery from the snowpack greatly influence the slope runoff contribution to the basins. This study focuses upon the comparative issue and addresses the spatial differences in the snowmelt hydrology of subarctic slopes.

Study Area and Methods

Two slopes in the lower Wolf Creek basin (61°31'40" N, 135°31'14"W), located 15 km south of Whitehorse, Yukon Territory, Canada, were selected for this study (Fig. 1). These slopes are separated by the Wolf Creek valley, which at this point is 120 m wide and at an elevation of 1,175 m above mean sea level. The north-facing (N) slope has a gradient of 0.25 and is underlain by boulder-clay (till), capped by an organic layer consisting of peat, lichens, mosses, sedges and grasses. Permafrost usually

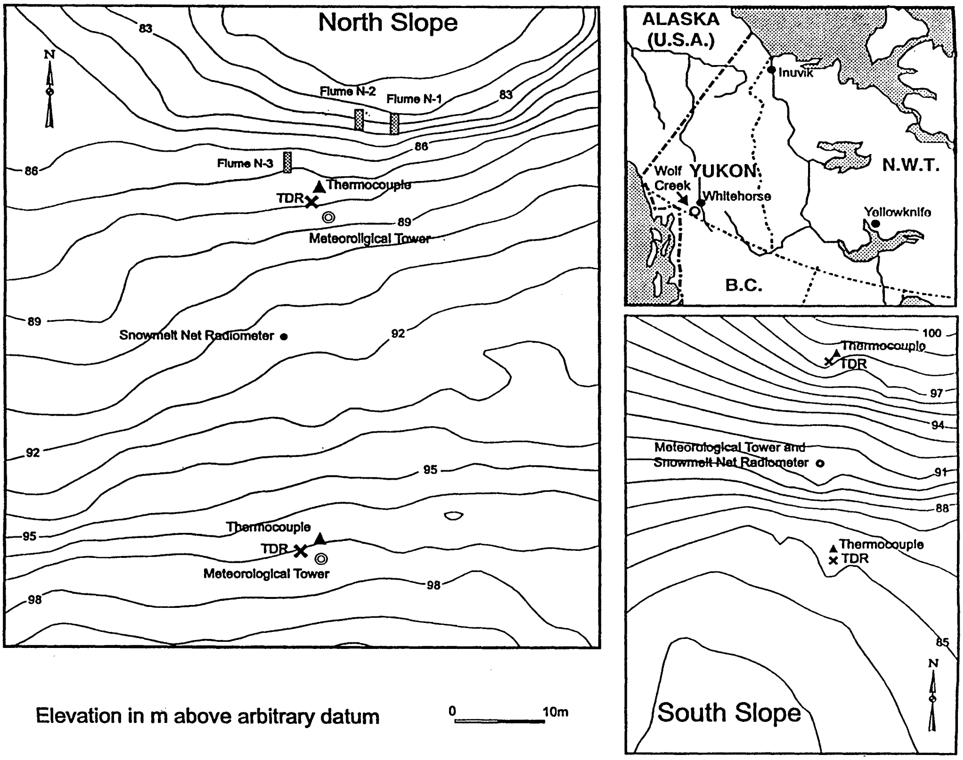

Fig. 1. Location of the study area (Wolf Creek basin), topography of the experimental slopes (North and South slopes) and the deployment of instrumentation.

occurs at depths of 0.6-1.2 m except at the slope bottom where seasonal frost depth exceeds 2 m. Most parts of this slope are covered by Labrador tea and willow shrubs, growing to a height of 1.6 m. Black spruce trees, approximately 20 m apart, reach heights of 9 m. The south-facing (S) slope has a gradient of 0.65. A leaf litter of 0.04 m overlies silty materials. No permafrost is found on this slope but winter frost penetrates to a depth of 1.5 m. A dense aspen forest, approximately 11 m high, covers the slope but the absence of leaves during the snowmelt period allows much radiation to reach below the canopy.

Wolf Creek basin has a subarctic continental climate. The long-term climatic record at Whitehorse yields a mean January temperature of -21°C and a July mean of 15°C. Temperatures during the snowmelt months of April and May average 1°C and 7°C respectively. Mean annual precipitation is 260 mm with about 55 per cent falling as rain.

This study was carried out from April 8 to May 28, 1997. Net radiation on each slope was measured by REBS net radiometers set at 1 m above the snow, air temperature and relative humidity were measured with a Vasiala HMP35CF sensor housed in a Gill shield. Ground temperatures were measured by two arrays of thermocouples on each slope, down to a maximum depth of 1.5 m. Soil moisture was determined by TDR probes buried at 0.05, 0.1, 0.2, 0.4 and 0.6 m in the N-slope, and at an additional depth of 1 m on the S-slope. Wind speed was measured by a Met One 014A anemometer set at 1.5 m above snow surface. These data were recorded by Campbell Scientific CR10 dataloggers. In addition, an eddy correlation device (Gill propeller anemometer Model 27106 and a Campbell Scientific FW3 fine wire thermocouple) was mounted at a 6 m height to provide information for the calculation of sensible heat flux over the N-slope. Eight snow pits were excavated in early April on each slope to determine snowpack thickness and density. Daily ablation was obtained by measuring the rate of snow surface lowering and converting the depth change into water equivalent using density measurements. During the snowmelt period, five flumes installed at the outlet and along three rills on the N-slope provided measurements of surface flow.

Snow Cover and Conditions During Melt

On April 8 when this study began, the snow was isothermal on the S-slope and some melting occurred despite daily mean air temperature being $<0^{o}C$. Mean snow depth on this date was 0.44 m which yields a water equivalent of 137±14 mm. The snow cover consisted of three visible layers (Fig. 2): a thin surface crust, a dense wet middle layer and a less dense layer that included some depth hoar. In contrast, temperature of the snow on the N-slope was $<0^{o}C$. Its mean snow depth was 0.71 m, representing a water equivalent of 187±24 mm (Fig. 2). Differences in end-of-winter snow storage on these slopes may be partly due to early melt on the S-slope, though a snow survey conducted in April 1996, before any ripening started, indicated that snow depth on the N-slope was 0.7 m compared to 0.5 m on the S-slope. A lower winter accumulation may be due to higher interception and sublimation losses from the S-slope.

Air temperature during the study period was similar for both slopes (Fig. 3). Daily means fell from near freezing to $-5^{o}C$ on April 18 and then rose sharply over the next five days, corresponding with rapid depletion of the snow on the S-slope. Afterwards, temperature remained around $+5^{o}C$ and then rose to $>10^{o}C$ when most of the snow on the N-slope also disappeared. Fig. 3 shows the positive values of daily net radiation (*i.e.* for the daylight hours only) for both slopes which increased rapidly when the air temperature rose above $0^{o}C$. Values for the S-slope exceeded that of the N-slope throughout the melt season, with larger differences occurring after the snow was gone from the S-slope. Wind speed was low along the sheltered valley during most of the study period. Only once did the daily average wind speed at 1.5 m above the snow exceeded 2 m s^{-1} (Fig. 3).

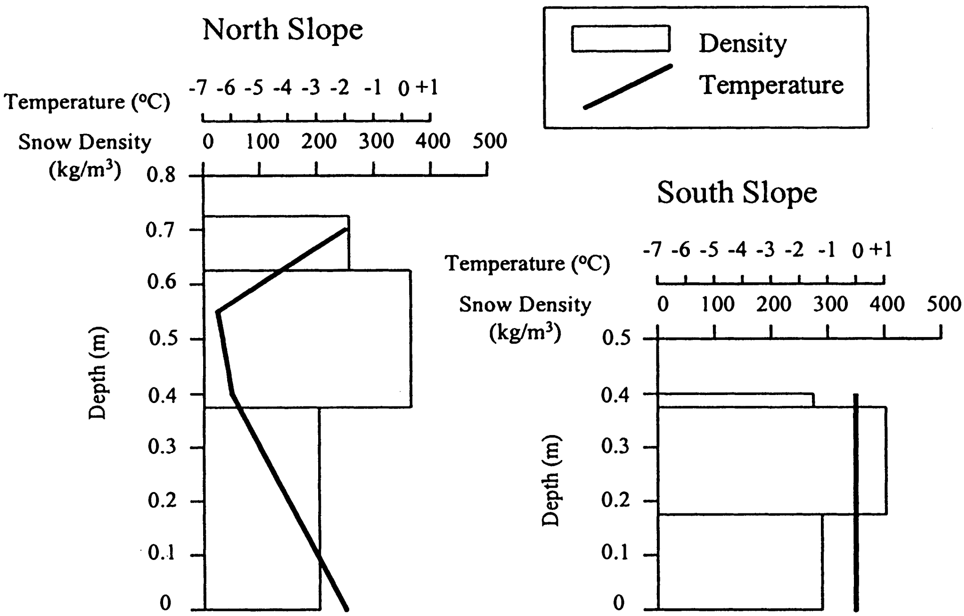

Fig. 2. Vertical variation of temperature and density in typical snow profiles on North and South slopes as measured on April 8 1997.

Snowmelt Processes

The energy balance of a melting snowpack is

$$Q_M = Q^* + Q_H + Q_E + Q_P + Q_S \tag{1}$$

where Q_M is the melt energy, Q^* net radiation, Q_H and Q_E sensible and latent heat, Q_P melt energy convected by rainfall, Q_S change in heat storage within the snow and all the terms are in MJ m^{-2}d^{-1}. In this study, Q_M was determined using ablation measurements

$$Q_M = L_f \rho M \tag{2}$$

where M (in m d^{-1}) is measured melt, L_f and r are the latent heat of fusion (MJ kg^{-1}) and the density of snow (kg m^{-3}). Only trace precipitation events occurred during the melt period and Q_P is ignored. The snow on the S-slope was already ripened and Q_S was unimportant, but on the N-slope, the cold content in the snow was calculated by

$$Q_S = c \sum_{i=1}^{3} d_i (T_i - T_0) \tag{3}$$

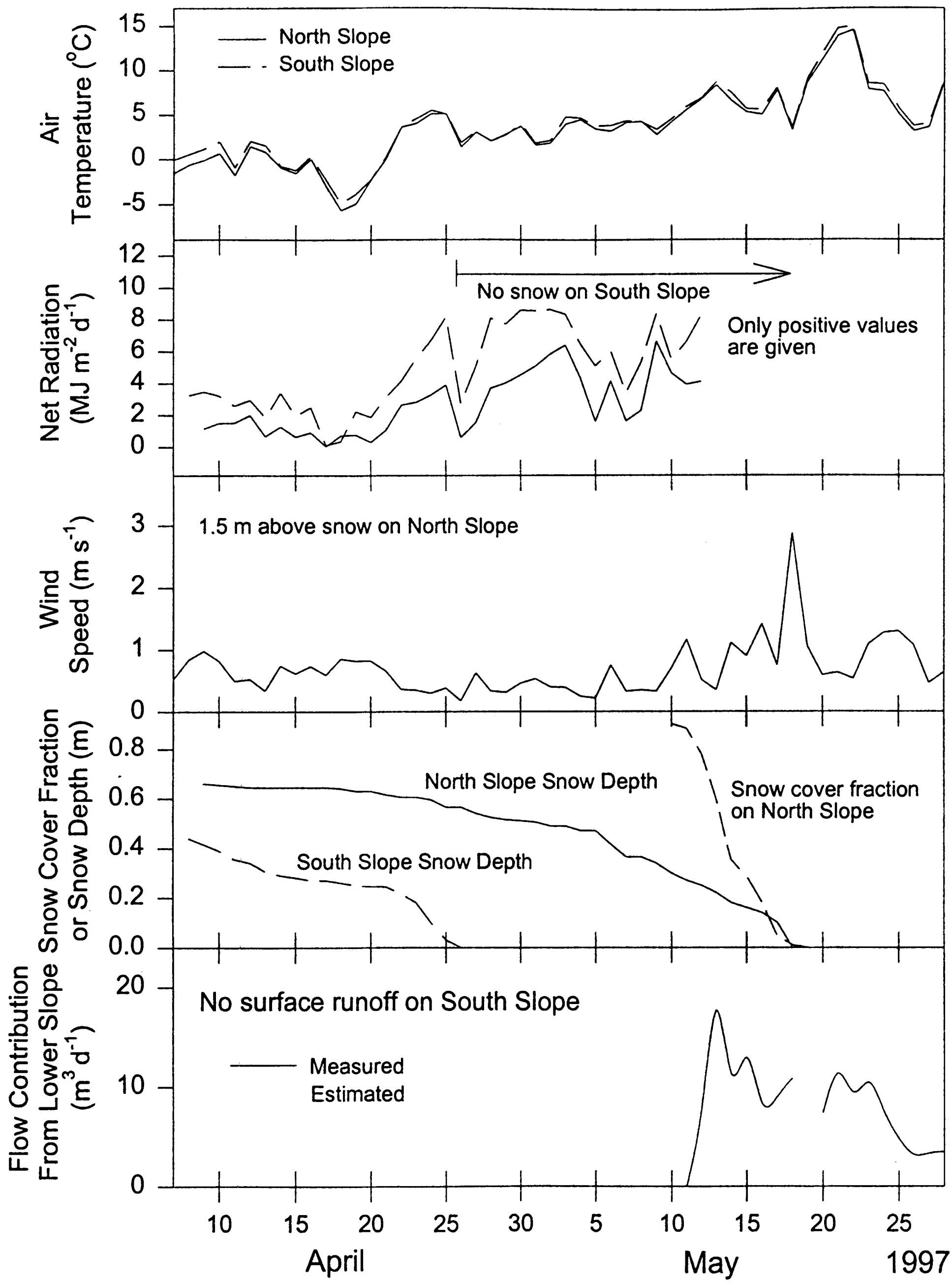

Fig. 3. Mean daily air temperature, daily net radiation (positive values only), mean wind speed, snow depth and snow cover fraction, and daily runoff contribution from lower North slope.

where c is the volumetric heat capacity of snow (MJ m^{-3}C^{-1}), T_o=0°C, d_i is snow thickness and T_i is snow temperature for layer i, for a total of three layers (Fig. 2). The low wind speed near the snow surface prevented calculations of Q_H and Q_E using aerodynamic equations. These terms were lumped as turbulent flux (Q_T), obtained as a residual of the energy balance

$$Q_T = Q_H + Q_E = Q_M - Q^* - Q_S \tag{4}$$

The magnitude of these energy balance terms was assessed (Table 1).

Table 1 – Snowmelt energy balance for North and South slopes, 1997

Energy fluxes (MJ)	North slope April 9-20	North slope Apr. 21-May 18	South slope April 8-24
Melt energy (Q_M)	+3.0	+74.3	+50.3
Net radiation (Q^*)	+11.3	+57.6	+45.1
Turbulent fluxes (Q_T)	-5.9	-16.7	+5.2
Heat storage change (Q_S)	+1.2	0	0

Fig. 3 shows the rates of snow cover depletion as represented by snow depth changes on the two slopes. The main melt period for the S-slope ended on April 26 except for several patches at upslope locations which lingered until May 8. The snow on the N-slope did not become isothermal until April 21 but after that, the main melt period continued until between May 10 to May 18 when the snow cover disintegrated into patches. Where there were icings in the hollows or along the rills, melting and erosion of the icings followed.

Net radiation was 45 MJ and Q_M was 50 MJ for the main melt period on the S-slope, leaving a net turbulent flux contribution of about 5 MJ. Such a dominance of radiation melt in a subarctic setting has also been reported elsewhere (Price and Dunne 1976). On the N-slope, Q^* was 11 MJ for the early melt period while Q_M was only 3 MJ. About 1 MJ was spent in eliminating the cold content, leaving a net Q_T of -6 MJ, with the negative sign indicating energy consumption by sublimation. In the main melt period, Q^*=58 MJ, Q_M=74 MJ and Q_S = 0 because the snow was isothermal. Turbulent flux was -17 MJ but it combined Q_H and Q_E which were likely to be opposite in flux direction. A rough assessment of Q_H was made for a large area of the slope using an eddy-correlation device placed at 6 m above the snow. Although the footprint of the sensors included some trees and protruding shrubs, the computed Q_H = 33 MJ was considered a reasonable approximation of sensible heat flux to the snow. Then, from Eq. (4), Q_E = -50 MJ which is equivalent to 18 mm of sublimation. This magnitude is similar to the atmospheric water loss of 21 mm reported by Kane *et al.* (1981), and 23 mm reported by Eaton and Wendler (1982) for the melt season on a north-facing slope near Fairbanks, Alaska.

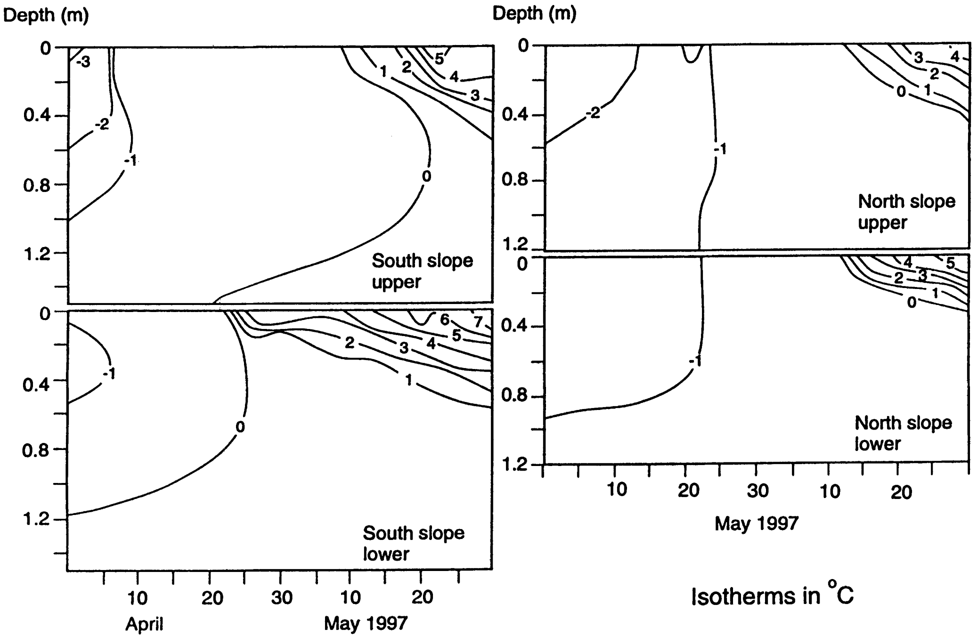

Fig. 4. Temperature changes in the near surface zones at four plots. For each plot, the disappearance of its snow cover corresponded with the rise of the surface temperature above 0°C.

Infiltration and Ground Temperature

In early April, ground temperature was <0°C at all sites except the lower S-slope which had seasonal frost extending down to 1.2 m only. The presence of a snow cover insulated the ground from conductive heating, but between April 5 and 8, ground temperature at the top 1-m layer of the upper S-slope rose abruptly by >2°C (Fig. 4). Similar rapid temperature rises were observed in the top layer of the N-slope sites but at later dates (April 22-24). This phenomenon may be attributed to the re-freezing of infiltrated meltwater, releasing heat to warm the soil (Woo and Heron 1981).

Temperature profiles of the S-slope indicate that only seasonal frost was present (Fig. 4). The 0°C isotherm could not descend from the surface while there was a snow cover, but ground thaw proceeded upward from the base of the seasonal frost. When the snow melted, two-sided thawing of the frost occurred. Ground frost was eliminated by April 25 at the lower slope, and the loss of frost at the upper slope was delayed until May 22 as the snow remained until May 8. Throughout the study period, no surface runoff was observed on the S-slope, suggesting unimpeded infiltration into the frozen soil (Kane and Stein 1983). The presence of seasonal frost on this slope had little noticeable effect on infiltration.

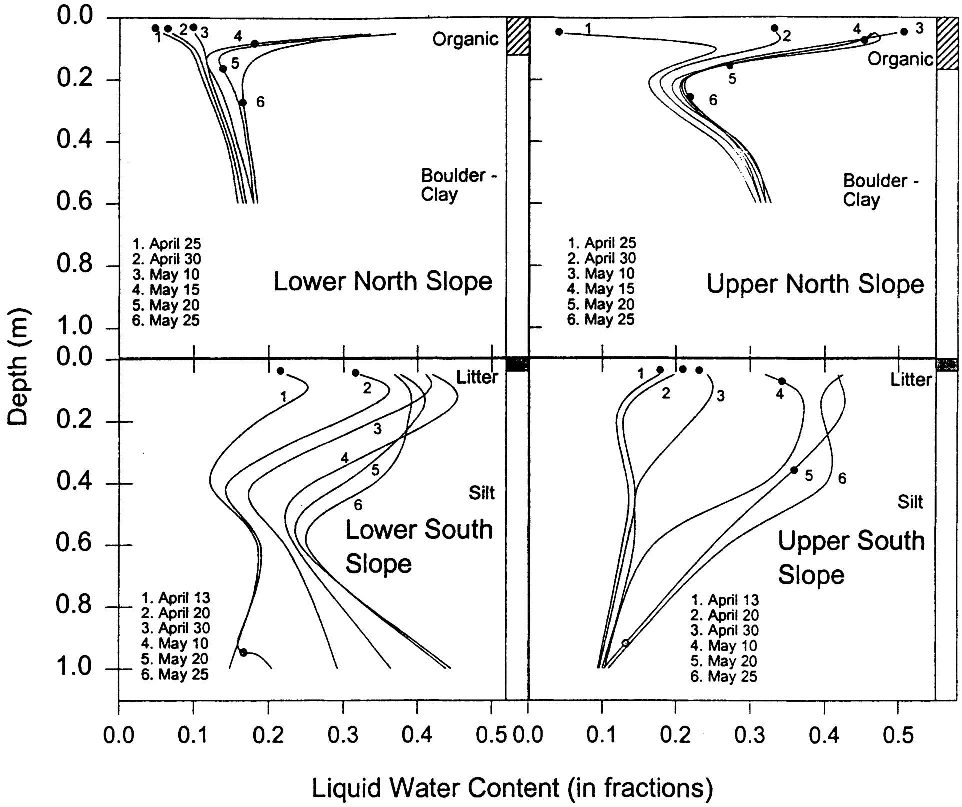

Fig. 5. Vertical profiles of liquid water content at four plots on six dates in the study period, 1997. Vertical columns on the right of each panel show the soil stratigraphy. Black circles indicate position of the downward thawing front; shaded circles for South slope profiles show the thaw boundary at the base of the seasonal frost. Profiles without any circle attached are frost-free.

Ground thaw on the N-slope began on May 12-13 with thawing proceeding downward only (Fig. 4), a feature typical of soils underlain by permafrost. By May 30, thaw depth reached only 0.35-0.4 m. The slow thaw may be attributed to the insulating properties of the organic cover and the abundance of ground ice which accentuates the 'zero curtain' effect, thereby retarding the descent of the thawing front (Woo and Xia 1996). Unlike the frost of the S-slope, the ice-rich zone at the organic-mineral interface limited meltwater percolation. This is evidenced by the continual saturation of the organic layer without a corresponding rise in the moisture content of the mineral soil below (Fig. 5).

Soil Moisture Changes

Liquid water content in the soil changes in response to 1) phase change as ground ice melts, 2) infiltration into the soil, and 3) lateral water movement. Fig. 5 shows selected profiles of liquid water content in the soils as measured by time domain reflectometry.

Between April 13 and 20, the lower site of S-slope showed an increase in liquid water content at the top 0.4 m layer. Since the soil was still frozen to a depth of 1 m, such an increase was due to infiltration of meltwater into frozen soil. After snow disappeared, the ground thawed quickly and further liquid water increase may be related to ground ice melt. This contribution to liquid water content was uniform throughout the profile, preserving the curvature established earlier in the season. After May 10, surface soil moisture decreased in response to evaporation and percolation to lower levels indicated by comparing the profiles of May 20 and 25. It is unclear whether the moisture increase at depth >0.8 m was caused by drainage from upslope.

On the upper S-slope, an increase in liquid water content near the ground surface between April 20 and 30 was likely due to infiltration into the frozen soil. Subsequent infiltration and percolation sustained a gradual moisture increase down to 0.4 m until May 10. After that, the snow was depleted; infiltration ceased but percolation continued to raise the moisture level down to a depth of 0.5 m. The moisture profiles at this site are similar to those presented by Kane and Stein (1983) for a non-permafrost slope in Fairbanks, Alaska.

Liquid water content of the organic layer on the lower N-slope did not increase notably in the melt period, suggesting a low infiltration rate which was a consequence of the considerable amount of ice that filled the pores in the organic layer. Thawing of the surface zone released much of the ice as liquid water (May 15). On the upper N-slope, the liquid water content of the top layer rose between April 25 and early May when the ground was still frozen. The ease of infiltration into frozen organics at this site is due to the large amount of air space partly created by the vapourization of the ice in the pores during the winter, a phenomenon documented elsewhere in the subarctic (Santeford 1979a; Smith and Burn 1987). Despite the high moisture content in the organic layer, percolation to the mineral substrate is hindered by an ice-rich zone at the interface of the organic and mineral soils, a feature also noted by Slaughter and Kane (1979). Consequently, once the storage deficit in the organic soil is satisfied, additional water input generates runoff (Santeford 1979b) which moves downslope as surface and subsurface flows.

Slope Runoff

At no time was the soil on the S-slope saturated and surface runoff did not occur. In contrast, many parts of the N-slope were saturated by the infiltrated meltwater and slope runoff was prevalent. When runoff started on May 12, much snow remained and water was issued at the base of the pack to topographic lows and the rills. As melt continued, water moved along preferential paths consisting of intermittent rills and pipes in the organic layer (*cf.* Gibson *et al.* 1993; Hinzman *et al.* 1993). Together, they formed a network which conveyed the flow rapidly downslope. In addition, subsurface diffused flow began when the saturated ground thawed, facilitated by the high hydraulic conductivity of the organic layer ($\approx 1 \times 10^{-4}$ ms^{-1}). Both the pathways and the source areas of runoff changed as new surface and subsurface hydrological connections were made. Snowpack location, micro-topography, water table position and thaw depth all played a role in altering the source area and the pattern of flow.

Flumes set up at the base of N-slope along three rills recorded snowmelt runoff between May 12 and 24 (Fig. 6). The areas draining into flumes are not constant as their source areas are altered by changes in the drainage network. In relative terms, flumes N-1 has the largest catchment area and N-2, the least. All flow data showed prominent diurnal cycles, with daily peaks occurring around 2130 h and daily lows at 0800 h (Pacific Standard Time). Runoff rose to high values on May 12 and then declined during the next three cool and overcast days. Increased temperature and radiation led to the second high flow period on May 18 but the flow at N-2 already dropped because of snow depletion in its catchment area. May 18 produced the highest flow at N-3 but N-1 peaked on May 21. The discordance in flow among the rills marked the differences in their water sources as the meltwater contribution area changed. As the snow cover on the slope disintegrated, the diurnal flow cycles disappeared and runoff at N-3 ceased on May 24. The recurrence of flow at N-2 on May 23 was not related to snowmelt but was due to replenishment by subsurface flow through a deepening thawed zone.

The timing of flow lagged behind the initiation of major snowmelt by at least two weeks (Fig. 3). Runoff began on May 12 when about half of the storage on N-slope and all the snow on S-slope was depleted. Such time lag is likely the result of meltwater movement and storage in the snowpack, the replenishment of water storage in the organic layer, seepage to the pipes and rills and transmission of water to the lower slope by surface and subsurface drainage. The highly variable source area and the changing pattern of the drainage network complicate the timing of meltwater delivery.

To assess the contribution of snowmelt to runoff, a water balance for the melt period is estimated. Since the total area draining into the three flumes cannot be determined accurately, two additional flumes were placed higher up the slope to catch the flow from upslope. The difference in the flows between the upper and the lower flumes (bottom of Fig. 3) is considered to represent the contribution from the area that drained between these flumes (inset of Fig. 6). Another difficulty is the missing data for the period from 1300 h, May 18 to 1500 h, May 20. A crude interpolation was made based on flow relationship with air temperature. Using a runoff contributing area of 840 m^2 (Fig. 6), the magnitudes of the water balance components for the

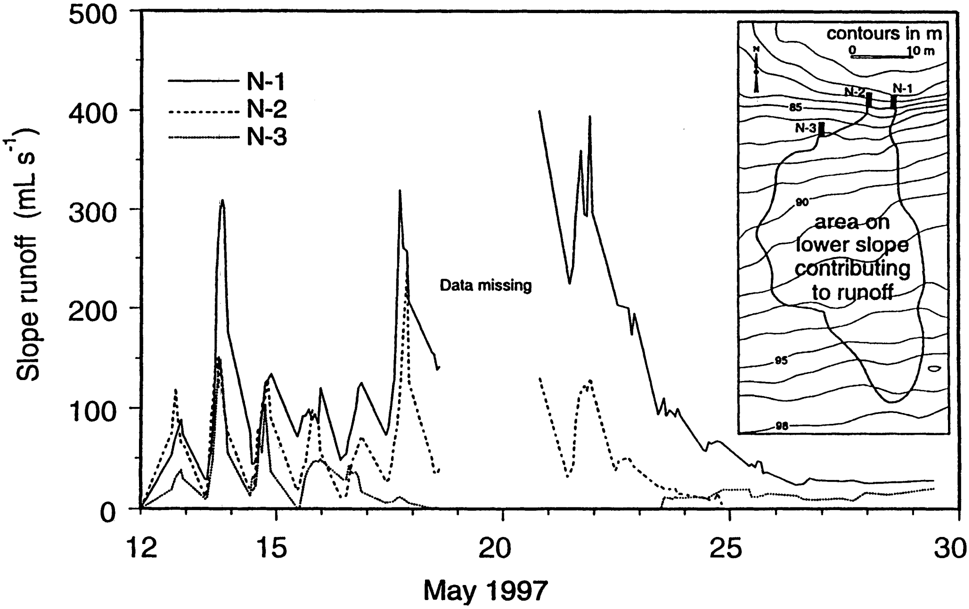

Fig. 6. Runoff at flumes N-1, N-2 and N-3 located in the rills at the base of North slope. Inset shows the slope area that contributed to runoff differences between the upper and the lower flumes.

melt season (April 21 to May 25) are calculated. The snow cover (187 mm water equivalent) lost 18 mm to sublimation, but icing on the slope provided 19 mm of water. Runoff was 155 mm, leaving 33 mm as water added to the active layer storage. The amount of meltwater that recharged the active layer is comparable to the 30 mm for a Fairbanks site (Santeford 1979b) and it is a plausible value in terms of change of moisture content measured in the organic zone (Fig. 5). The runoff ratio, being 155/(187-18+19), was 0.8. This value is higher than the ratio (0.5) obtained by Kane *et al.* (1981) for an experimental plot near Fairbanks where the snowfall is substantially less; but it falls within the range of snowmelt runoff ratios (0.6-0.8) obtained from runoff plots set up near Yellowknife, Northwest Territories, Canada (Landals and Gill 1972).

Discussions and Conclusions

The selection of two subarctic slopes close to each other eliminates meso-scale variabilities so that differences in their hydrological behaviour can be attributed to local considerations alone. Slope aspect and its attendent physical attributes such as incoming radiation, vegetation, frost and soil development, lead to large contrasts in snowmelt and meltwater delivery. Results from one field season clearly demonstrate large differences between a north and a south facing slope in relation to the magnitude, timing and the mechanisms of melt and runoff processes.

There are several major contrasting hydrological features between the two slopes.

1) At the end-of-winter, there is less snow on the south than the north slope and this seems to be a function of interception and sublimation loss over the winter or the pre-melt period. By the time snowmelt begins in earnest on the north slope, the snow has gone from the opposite slope even though the latter has a much higher density (though leafless) of trees.
2) The snow on the south slope ripens more than 10 days earlier than the north slope. Radiation melt is important and this may be one factor that accentuates the contrasts in the timing and the magnitude of snowmelt between slopes with different amounts of radiation receipt.
3) Seasonal frost is found in the south slope but this is no deterrent to meltwater infiltration. The frost, >1 m thick, thaws out quickly at the end of snowmelt.
4) Snowmelt does not generate direct runoff on the south slope. However, runoff may occur in certain years when pre-melt soil moisture conditions are high, as was observed in a similar environment near Fairbanks, Alaska (Slaughter, pers. comm).
5) The north slope is underlain by permafrost with a thin active layer, insulated by an organic soil with living moss and lichen. Ground thaw below the organic zone is retarded by an ice-rich layer at the organic-mineral soil interface.
6) Meltwater infiltration into frozen organic soil depends on the ice content in the soil pores. Percolation to the mineral soil is limited by frost and runoff is generated when the organic layer reaches saturation.
7) Delivery of water downslope is facilitated by piping in the organic layer and by rills that channel the flow. The source area and the drainage network for slope runoff are highly dynamic, changing according to the snow disposition and the convergence or divergence of surface and subsurface flows.
8) There is a considerable time lag (>10 days) between the initiation of snowmelt and the occurrence of runoff on the north slope because of the various storages (snow, organic soil) and delays in water transmission (water movement in snow, surface and subsurface flows).

From the perspective of subarctic catchment hydrology, it is important to understand the mechanisms of meltwater production and delivery, and to quantify the timing, magnitude and spatial distribution of slope runoff. Characteristics such as those manifested by the south slope will dominate catchments with only seasonal frost while hydrologic features present on the north slope will prevail in basins with considerable permafrost. Thus, the former type of basins yields little runoff in the melt period but the latter group will have flashy flows. This is verified by Slaughter *et al.*'s (1983) comparison of streamflow for three headwater basins in the Caribou-

Poker Creeks watershed, Central Alaska, where the catchments with larger percentages of permafrost have higher and flashier discharges. In terms of snowmelt modelling for small catchments, particularly for distributed models (Kirnbauer *et al.* 1994), proper account should be taken of the spatial contrasts in slope runoff production in order to generate streamflow that reflects the variable source areas. Macro-hydrological models which lump large areas into single units should not ignore the intra-unit variability in the timing, the area and the magnitude of meltwater production as demonstrated in this study. This, however, involves scaling considerations (Blöschl and Sivapalan 1995) which cannot be easily resolved.

Acknowledgements

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