Subsurface Drainage from Organic Soils in Permafrost Terrain: The Major Factors to be Represented in a Runoff Model

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ABSTRACT: Field studies on subsurface runoff from organic terrains underlain by permafrost are underway at different site types in northwestern Canada. The objective of these studies is to identify the important factors controlling the timing and magnitude of subsurface drainage, in order to assist the development of algorithms for runoff modelling for this environment. Initial results suggest that organic-covered permafrost terrains share several important hydrological characteristics. At each site, the hydraulic conductivity in the organic soil decreased exponentially with depth. It is therefore necessary to know the elevation of the saturated layer in order to estimate the rate of lateral drainage. The elevation of this layer is limited by the relatively impermeable frost table, which lowers with time during ground thawing. A strong correlation between cumulative degree-day ground surface temperature and the portion of the cumulative ground heat flux used to melt ice in the active layer is demonstrated. The water table responds rapidly to hydraulic inputs to the ground surface, due to very high infiltration rates, the close proximity of the saturated zone to the ground surface, and to relatively low specific yield of well decomposed peat. The near-surface soils were found to have two seasonally-characteristic moisture regimes, a wet regime during snowmelt runoff, and a dry regime beginning shortly following this period. It is suggested that the absolute values of each regime vary among the study sites because of differences in local hydraulic gradient.

1 INTRODUCTION

In Canada, continuous organic terrains underlain by permafrost include the tundra, taiga and high-boreal regions, which collectively occupy approximately one-third of the Canadian land surface (Bliss and Matveyeva, 1992). In these areas the organic soil is generally 0.2 to 0.5 m thick (Slaughter and Kane, 1979) and consists of a 0.1 to 0.25 m layer of living and lightly decomposed vegetation, overlying a more decomposed layer (Hinzman et al., 1993). In most cases, the water table remains within the organic soil throughout the thaw period (Quinton and Marsh, 1999). Runoff processes in these organic-covered terrains are distinct from those of other permafrost areas. For example, many organic-covered terrains in cold reions experience little or no surface flow, due to the large water holding capacity of organic soils and their high frozen and unfrozen infiltration rates (Dingman, 1973), which far exceed the rate of input from snowmelt or rainfall. Melt water can percolate through the unsaturated, highly porous organic soil and move rapidly downslope, even while the ground is still snow-covered (Quinton and Marsh, 1999), and before the moisture deficit of the soil is satisfied (Chacho and Bredthauer, 1983).

In modelling the amount and timing of runoff water reaching stream channels, the processes controlling lateral flow must be properly represented. Quinton et al. (2000) demonstrated that lateral flow through the organic soil is laminar and can be described by the Darcy-Weisbach expression, $f = C/N_R$, where f is the friction factor, C is a coefficient and N_R is the Reynolds number. The authors suggested that a first approximation for a model of the flow regime may consider the organic soil as a single, continuous layer with depth-varying, flow resistance properties. Quinton and Gray (2001) suggested a physically-based approach of estimating the cumulative flux of energy into the ground as a first step toward estimating the change in frost table elevation during soil thawing.

This paper identifies and discusses the factors controlling the timing and magnitude of subsurface drainage, in order to assist the development of algorithms for runoff modelling in this environment. This will be accomplished through the presentation of field measurements of soil transmission properties, thermal development, soil moisture storage and water table response.

2 STUDY SITES

Field studies were conducted at three northern locations, each representing a widely-occurring biophysical land cover type of cold regions: Granger Creek (60° 31'N, 135° 07'W), a sub-arctic, alpine catchment, 15 km south of Whitehorse, Yukon; Scotty Creek (61° 18'N, 121° 18'W), a wetland-dominated, high boreal catchment, 50 km south of Fort Simpson, NT; and Siksik Creek (68° 44'N, 133° 28'W), located at the northern fringe of the forest-tundra transition (Bliss and Matveyeva, 1992), 55 km north of Inuvik, NT in the arctic tundra. Permafrost is present at all three sites, and each site has a continuous organic cover, with the thickest accumulation at Scotty Creek (~0.5 to 0.8 m), followed by Siksik (~0.2 to 0.5 m), and then Granger (~0.2 to 0.35 m). By late summer, the average frost table depth is ~0.6 m at Scotty and Granger, and 0.3 m at Siksik.

3 METHODOLOGY

At each site, soil cores were extracted that contained the entire thickness of the peat layer. The cores were divided in half length-wise and on one half, three to five sub-samples were taken at evenly-spaced intervals in order to measure the bulk density and total porosity. The other half were sent to the Department of Land Resource Science, University of Guelph, for the development of soil thin sections. These sections were then used for the measurement of active porosity and mean pore diameters.

At Siksik Creek, measurements of the depth of the frost table below the ground surface were made weekly along a transect at fixed distances from the edge of a stream using a graduated steel rod. Frost table depth was measured less frequently at Scotty and Granger, since these sites were not continually staffed. These measurements were made and recorded hourly. At each site, thermistors and time domain reflectometer probes were deployed in a vertical profile in a soil pit in order to monitor temperature and volumetric liquid water content at regular intervals. At Siksik Creek, the spacing of these sensors corresponded to the depth at which sub-samples were taken from the soil cores for the measurement of physical properties. At the other sites, where the frost table depth was not regularly measured, the sensors were spaced more closely so that the depth of the frost table could be monitored as the soil thawed.

The volumetric heat capacity for the position of each sensor was computed from the product of the bulk density and the specific heat, with values of the latter obtained from the literature (Table 1). The soil profiles were divided into soil layers based on the spacing of the sensors, with each sensor location representing the middle of a soil layer. The composite volumetric heat capacity of each soil layer was then computed from the sum of the volumetric heat capacities and fractional volumes of the soil, air, ice and water constituents, using the expressions provided by Quinton and Gray (2001). Changes in the fractional volumes of ice, water and air were monitored from the measurements of soil moisture, frost table depth and soil temperature. In addition, the elevation of the water table at the soil profiles, was monitored using a float recorder mounted on a stilling well. The thermal conductivity of the soil layers, 8 was computed using the equations provided by Farouki (1981). The heat flux into the ground Q_G, was then derived from Fuchs (1986):

$$Q_G = Q_A + Q_P + Q_M \quad (1)$$

where Q_A is the energy used to warm the active layer, Q_P is the energy used to warm the permafrost, and Q_M is the energy used to melt ice in the active layer. Q_G was also directly measured, along with the ground surface temperature (T_S) at the meteorological towers of Environment Canada at Siksik and Granger Creek, using a ground heat flux plate inserted at 4 cm depth, and an thermal infra-red thermacouple. However, because both stations, were operable only intermittently during the study periods, they provided only limited data for this study.

Table 1. Volumetric composition of soils, and thermal properties used for computations.

layer		total por.	spec. yield **	specific heat †	heat capacity *	thermal cond.
top	bottom	0_T	Sy	c	Cv	8
m	m	-	-	$Jkg^{-1}K^{-1}$	$Jm^{-3}K^{-1}$	$Wm^{-1}K^{-1}$
0.00	0.05	-	35	-	-	-
0.05	0.15	0.96	30	1920	78912	0.21
0.15	0.25	0.90	20	1920	144384	0.21
0.25	0.35	0.87	10	1920	175392	0.21
air	-	-	-	1010	1212	0.025
ice	-	-	-	2120	1950400	2.24
water	-	-	-	4185	4185000	0.57

*heat capacity of the soil phase,†Values obtained from Miller (1981) and Oke (1987),**measured at Scotty

Horizontal flow rates through the peat soils were measured from chloride tracer tests at Siksik Creek in 1993 and 1994, at Scotty Creek in 1999 and 2000, and at Granger Creek in 2000. From these tests, the average pore velocity v, was computed as,

$$V = L_X \ / \ t_C \quad (2)$$

where L_X is the distance between the tracer application line and tracer-sensing electrode, and t_C is the length of time between the application of the tracer and the time when the centre of mass of the tracer plume reached the sensing location. The hydraulic conductivity (k) was then computed from:

$$k = v \times 0_A \times (dl/dH) \quad (3)$$

where 0_A is the active porosity measured from the soil cores, dl is the length between the points where tracer was applied and where it was measured, and dH is the difference in hydraulic head between these two points. For a detailed description of the tracer experiments the reader is referred to Quinton and Marsh (1999).

Vertical flow rates through the peat soil were measured from falling-head experiments at Siksik Creek using ring infiltrometers. Multiple runs were conducted over a variety of ground surfaces in order to obtain the average rate that water infiltrates the ground. Rainfall was also measured with a tipping bucket gauge at each study site.

4 RESULTS

Since the peat soil is increasingly decomposed with depth, other physical properties also vary with depth. The total porosity shows a slight decrease with depth (Figure 1), while the active (i.e. inter-particle) porosity decreases relatively sharply from approximately 0.8 near the surface to between 0.5 and 0.6 at depth. This suggests that as depth increases, the proportion of pores that store water (e.g. closed and dead-end pores in plant tissues) increases, while the proportion of pores that transmit water decreases. This factor, in addition to the decrease in the geometric mean diameter of inter-particle pores with depth (Quinton et al., 2000), contributes to the decrease in the hydraulic conductivity with depth observed at all sites. Figure 2 shows that the hydraulic conductivity can decrease by as much as 5 orders between the upper and lower boundaries of the peat cover. Although the peat thickness varied among the study sites, the hydraulic conductivities at both upper and lower ends of the transition were of similar magnitude, suggesting that the shallow peat accumulations, such as at Siksik Creek, have a more abrupt decrease in hydraulic conductivity with depth. An association between depth and hydraulic conductivity is therefore needed in order to estimate the drainage rate for any saturated layer elevation. Using data from Siksik Creek, Quinton and Gray (2001) suggest using the linear association between depth and a coefficient C that relates the friction factor *f* and the Reynold's number, N_R (i.e. $f = C/N_R$).

An association between depth and hydraulic conductivity in itself is not sufficient to estimate the rate of subsurface drainage, since selecting a realistic value of k requires that the elevation of the saturated layer be known. The lower elevation of this layer is the relatively impermeable frost table, the depth of which increases with time during soil thawing. Prediction of the depth to the frost table would serve as a first approximation for the elevation of the saturated layer. A strong association between the cumulative mean daily ground surface temperature (ETs) and the cumulative daily Q_G was observed at all sites. Figure 3 shows this association using both computed (equation 1) and measured Q_G for the 30 day period of surface temperature measurement at Siksik Creek.

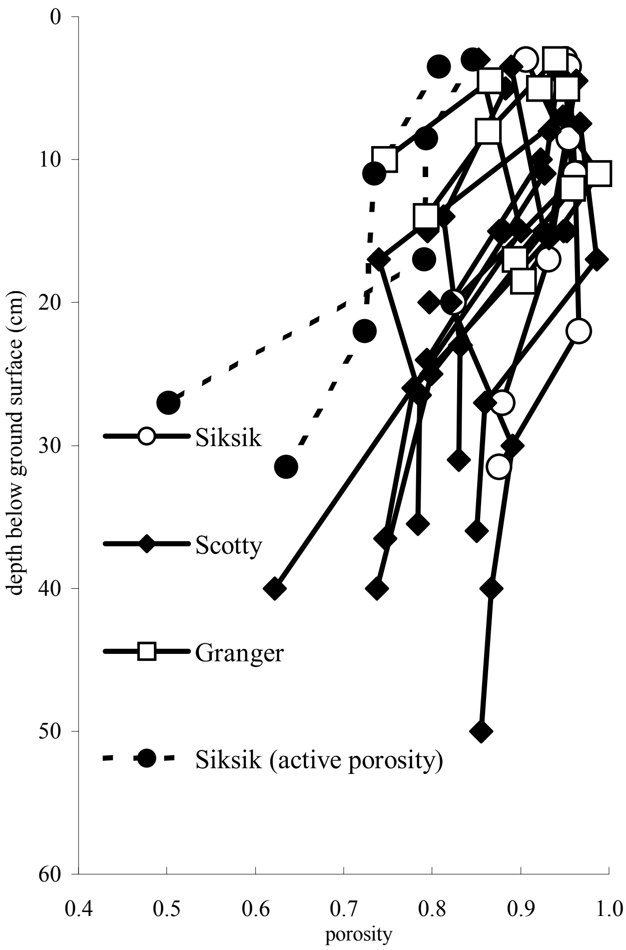

Figure 1. Variation in total (solid lines) and active (dashed lines) porosity with depth at the study sites.

Since segregated ice was not observed in the active layer above the frost table, it is reasonable to assume that Q_M is largely consumed at the bottom of the thawed layer. Therefore, if the value of the cumulative amount of energy consumed by Q_M is known, the corresponding drop in the elevation of the frost table m, can be computed from:

$$m = (Q_M / 8\ Di) fi \qquad (4)$$

where Q_M is in units of Joules, 8 is the latent heat of fushion, Di is the density of ice, and *fi* is the volume fraction of ice at the frost table. It is assumed that at the frost table, *fi* = 0_T (i.e. the soil is assumed to be saturated with ice). Figure 3 also shows a strong association between ETs and the cumulative daily Q_M (R^2=0.98), suggesting that the if the initial frost table position is known (e.g. at the time that the ground surface becomes free of snow), and if the cumulative mean daily ground surface temperature since that time is also known, then the depth to the frost table is obtainable by equation 4.

The final infiltration rates of 12 falling head experiments ranged between 24 and 300 mm min^{-1} (mean = 103 mm min^{-1}). Since the water table is usually within 50 cm of the ground surface throughout the year at all sites, these findings suggest that water infiltrating the ground surface can arrive at the water table within a few minutes. The high infiltration rates would account for why the water table began to rise within the same hourly measurement period as the start of the rain event (Figure 4). In this example, 23 mm of water was added to the ground surface in a single rain event at Scotty Creek. Prior to this input, the water table was 43 cm below the ground surface. At this depth, the specific yield is typically 10 and 20% (Table 1), and as a result, this depth of rainfall input generated a water table rise of approximately 15 cm (figure 4). As shown in Figure 2, a water table rise of this magnitude can increase in the hydraulic conductivity of the saturated layer by two or more orders.

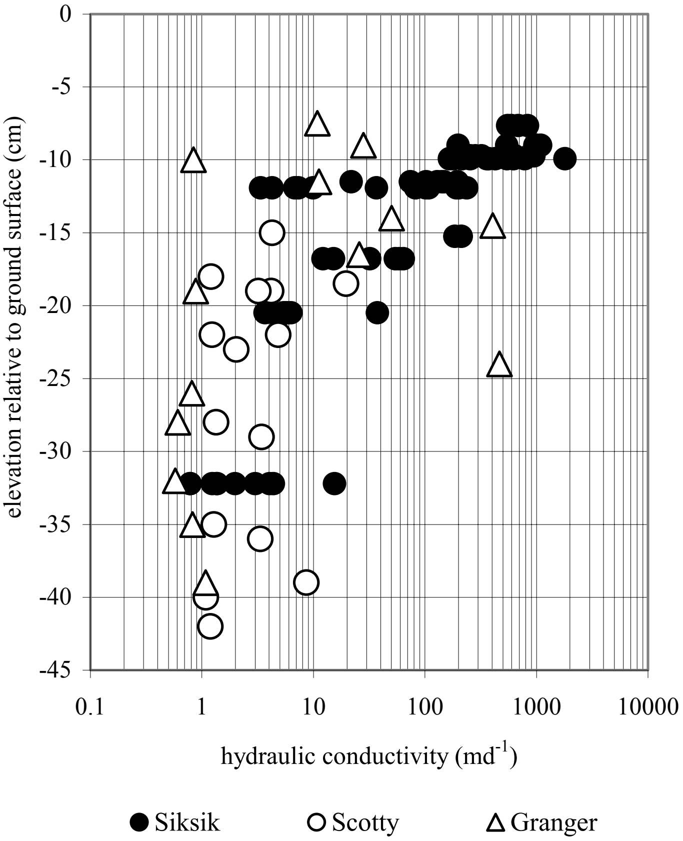

Figure 2. Variation in hydraulic conductivity with depth at the study sites.

Initial field measurements suggest that the soils at the study sites have two seasonally-characteristic moisture regimes. During spring, when the relatively impermeable frost table is close to the ground surface, and the snow melt water supply is large, the soil is in a wet regime. As the soil drains and thaws, the soil moisture adjusts to the lower value of the dry regime. The variability of soil moisture during each regime is small, and the transition time from the wet to the dry regime was less than 5 days at Granger and Siksik.

Table 2 compares the volumetric soil moisture 2, values of both regimes at each site for the 0-10 cm layer. At Siksik Creek, the measurements were made in a relatively flat (gradient = 0.08) area at the foot of a hillslope, and as a result, during the wet regime, 2 approached saturation during the wet regime, i.e. 2 . 0_T. A large moisture supply was maintained at this site, owing to the presence of a late-lying drift upslope. Following the melt runoff period, soil moisture reduced to an average value of approximately 0.28, which characterised the dry soil regime at this site.

Table 2: Mean and standard deviation of hourly soil moisture measurements for wet and dry moisture regimes at each study site

	wet regime		dry regime	
site	mean	st.dev.	mean	st. dev.
Siksik	0.93	0.015	0.28	0.015
Scotty	-	-	0.451	0.013
Granger	0.537	0.057	0.193	0.014

At Granger basin, the soil was below a melting snow cover during the first five days of measurement, and for the next 23 days, continued to receive meltwater drainage from a late-lying snow drift upslope. Although both the Granger and Siksik Creek sites were effected by a late-lying snow drift, unlike at Siksik, the upper limit of the volumetric soil moisture at Granger remained well below the value of the total porosity. It is suggested that the reason for this difference is due to the higher hillslope gradient (0.255) at Granger. The difference between the maximum soil moisture values 2_{max} of the wet regimes of Siksik (0.95) and Granger (0.64) suggests the relation:

$$2_{max} = 0_T - (0_A \times dH/dl) \qquad (5)$$

where 0_T and 0_A are the total and active porosities respectively, and dH/dl is the hydraulic gradient. Using the values of $0_T = 0.9$ and $0_A = 0.8$, representative of both sites, equation 5 estimates a maximum volumetric soil moisture of 0.70 for Granger and 0.84 for Siksik, both of which agree closely with the measured maximum values. As at Siksik Creek, once the late-lying snow drift had melted, the soil moisture at Granger underwent a brief transition period of approximately 4 days before adjusting to the dry moisture regime.

Soil moisture measurements at Scotty Creek began relatively late (21 June) and as a result, the values in the 0-10 cm zone represent the dry soil

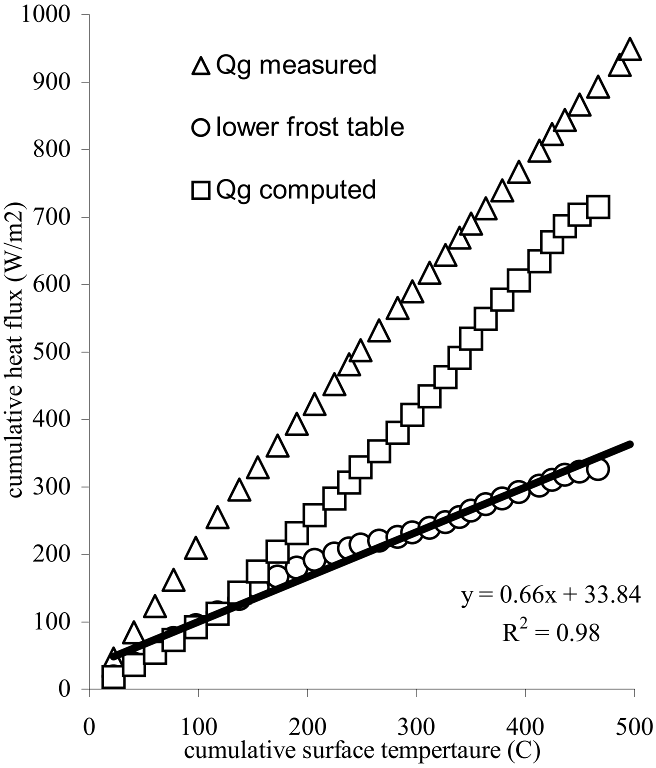

Figure 3. Relationships between the cumulative daily ground surface temperature and the ground heat flux (measured and calculated) and the component of this flux used to lower the frost table. For the latter, a linear association is defined.

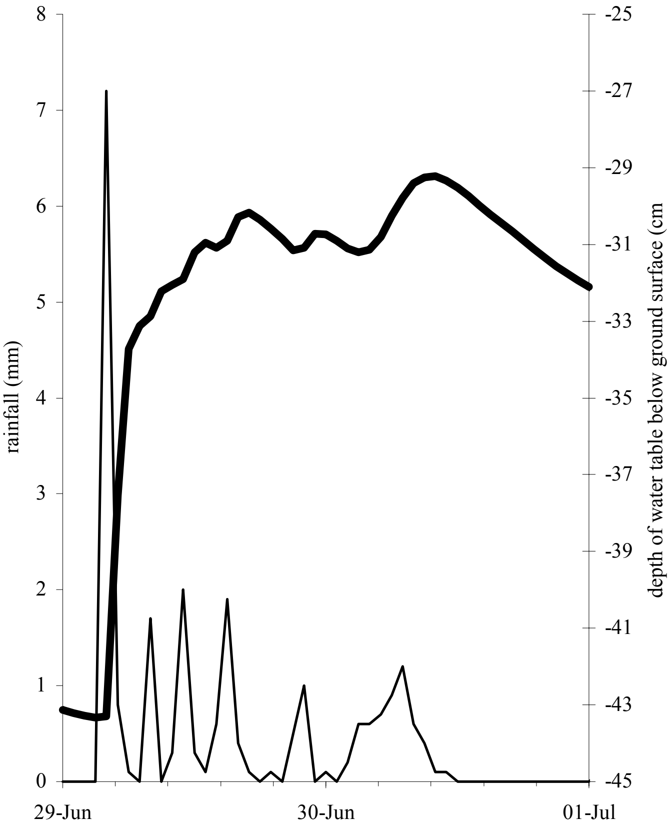

Figure 4. Response of the water table to a 23 mm rainfall event at Scotty Creek, 29 June, 1999

moisture regime only. Over the three-month period of record, soil moisture varied over a relatively narrow range between 0.43 and 0.49, and therefore had a low standard deviation as at the other sites (Table 2). The mean soil moisture at this site is relatively high given the nearly flat surface, and constant high water supply typical of a wetland environment. The soil moisture at this site, in combination with the water level and rainfall data, gives valuable insight into the water transmission and storage processes within the peat. For example, in response to the 23 mm rainfall event shown in Figure 3, the soil moisture in the 0-10 cm increased only from 0.46 to 0.49, suggesting that while some re-wetting of the soil occurred, most of the infiltrated rain water was transmitted (through the active porosity) to the saturated layer.

5 DISCUSSION AND SUMMARY

Several key hydrological properties of the organic soils of permafrost terrains have been presented. This will help to direct the development of physically-based runoff models for this environment. Correct delineation of subsurface lateral flow requires the definition the association between depth and hydraulic conductivity. Because the hydraulic conductivity appears to change more abruptly with depth at sites with thinner peat covers, it is likely necessary to define site-specific depth-conductivity associations, or to investigate the depth-conductivity association as a function of peat thickness. Since the frost table is relatively impermeable, the elevation of the saturated layer depends on the degree of soil thaw. An algorithm of mass flow must therefore be coupled to a heat flow routine in order to estimate subsurface runoff from these hillslopes during soil thawing. The strong correlation between the cumulative average daily ground surface temperature ΣT_s and proportion of the cumulative average daily ground heat flux used to lower the frost table ΣQ_M, suggests that the frost table depth can be computed. The association between ΣT_s and ΣQ_M also offers the prospect of using remotely-sensed thermal infrared imagery to estimate the rate of frost table lowering with time.

The very high measured infiltration rates, in combination with the observation that the water table is almost always within 50 cm of the ground surface, suggests that water applied to the ground surface is transmitted to the water table relatively quickly. As a result, the water table begins to rise very soon after water reaches the ground surface, and the magnitude of this rise can be computed if the input depth and the specific yield are known. Field measurements of the specific yield show that this property decreases with depth. Therefore estimation of the height of the water table rise in response to input requires that a depth-specific yield association be defined. When the water table is near the bottom of the peat accumulation, the specific yield can approach 10%. In this case, not only would the water table response be rapid, but it would also be relatively large in magnitude. Given the exponential change in hydraulic conductivity with depth, a rapid and large water table rise, could potentially produce a rapid and large runoff response from hillsides and stream channels.

Soil moisture is an important variable as it affects the soil thermal properties, and therefore the amount of energy available to lower the frost table, which controls the hydraulic conductivity of the saturated layer, and ultimately the rate of subsurface drainage. The soils at the study sites were found two have two distinct, seasonally characteristic moisture regimes: a wet regime when the snowmelt moisture supply is large, and the ability of the soils to store water is limited by the close proximity of the relatively impermeable frost table to the ground surface. The variability of soil moisture within each regime was low, and the adjustment from the wet to the dry regime at the end of the snow melt runoff period was rapid. The rapid process of water delivery from ground surface to the water table described above, was found to result in only a minimal change in soil moisture in the unsaturated zone.

Although this general pattern occurred at all sites, the absolute moisture values of the regimes varied among the sites. Field observations suggest that this variability is largely controlled by the local slope gradient at each site. At Siksik Creek, where the ground surface is relatively flat, the maximum soil moisture of the 0-10 cm layer during the wet regime approach saturation, while at Granger, where the ground surface has a relatively large slope, the maximum soil moisture is considerable below saturation. The active porosity has the ability to transmit water, but the rate of transmission is proportional to the hydraulic gradient. Where there is no gradient, the maximum soil moisture would be limited by the total porosity. It is suggested that the maximum soil moisture would decrease with increasing gradient until the lower limit is reached, i.e. when there is no water remaining in the active porosity, and $2_{max} = 0_T - 0_A$.

In addition to changes in the thermal and insulating properties of the near-surface soils, the adjustment from the wet to the dry soil moisture regime would have other important hydrological implications, not considered in the present study. For example, the stomatal resistance would abruptly increase when the soil enters the dry regime, and as a result, vertical moisture losses to evaporation would decrease.

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