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A Conceptual Framework for Runoff Generation in a Permafrost Environment

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Abstract:

Hillslope runoff processes were studied in the tundra region of the Canadian western arctic in order to provide a physically-based, conceptual framework for runoff generation for basins in this environment. It was found that subsurface flow is the dominant mechanism of runoff to the stream channel, that this flow is conveyed predominantly through the peat of the inter-hummock channels, and that subsurface flow through the highly conductive upper peat layer and soil pipes can be as rapid as surface flow. Stream discharge was computed from computations of hillslope runoff and meltwater input from the snowpack in the stream channel. The similarity between the computed and measured stream discharge suggests that the approach used to compute discharge would be useful as a conceptual basis for a distributed hydrological model for this environment. Hydrological interaction among the basin subsystems was found to play an important role in runoff generation. For example, when the water table was in the highly conductive upper peat layer, and discharge from inter-hummock channels was relatively large ($\sim$0·1–1·0 m^3 d^{-1}), the near-stream area, a zone of relatively thick peat, offered little attenuation to water draining toward the stream channel, and the source area for stream flow was relatively large. The duration of this high flow regime varied among hillsides due to variations in the width of the near-stream area, the gradient, length and inclination of hillslopes, and the water equivalent and melt rate of the snowpack. When the water table subsided into the lower peat layer, and discharge from inter-hummock channels decreased ($<$0·1 m^3 d^{-1}), the runoff source area was relatively small. Under this condition, if drainage into the near-stream area increased, the lateral expansion of the runoff source area would be delayed due to the relatively large storage capacity near the stream. Copyright © 1999 John Wiley & Sons, Ltd.

KEY WORDS hillslope runoff; stream flow; arctic; tundra; peat

INTRODUCTION

In recent years, the need to understand and predict the impact of persistent and complex perturbations, such as a warming climate, on the hydrology of natural systems has increased (WCRP, 1994; IPCC, 1996). This is especially true for permafrost environments which are sensitive to changes in energy and water availability, and as a result there is a need for physically-based hydrological models applicable to these environments. The robustness of such models would be enhanced if they were developed from a conceptual framework based on the physics of the dominant runoff processes.

Much of the literature on runoff processes and pathways in permafrost areas utilizes theories developed for temperate climates that are then adapted in order to explain the volume and timing of storm flow events

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(whether these are rainstorm or snowmelt driven), where the term storm flow has a time-based definition, and refers to water arriving at the stream channel in time to contribute to the hydrograph response (Chorely, 1979). An early example of this type of research is Dingman (1973) who modified the variable source area concept of Hewlett and Hibbert (1967) to describe runoff in the subarctic, boreal forest zone of Alaska. Dingman (1973) suggested that the very high vertical permeability of the ground cover in this environment makes overland flow extremely unlikely, except in valley bottom areas, where rapid runoff due to overland flow originates from a source area, the size of which depends upon antecedent soil moisture conditions. In this environment, storm flow originates in a small portion of the valley bottom, with the source area varying in size.

Numerous studies since Dingman (1973) have demonstrated that the Arctic is not a uniform region where a single runoff mechanism dominates. In many areas, for example, overland flow is extremely important for reasons other than those described by Dingman (1973), including large snowmelt inputs and low infiltration rates of the frozen mineral soils. Examples are provided by Marsh and Woo (1981), Woo and Steer (1982) and Lewkowicz and Young (1990) who showed that for areas of the Canadian high arctic, the runoff regime is dominated by the occurrence of late-lying or perennial snow drifts, which in turn are controlled by a combination of atmospheric and land form factors (Sturm *et al.*, 1998). As these patches melt over the summer, they have two important effects. First, meltwater released by the snow patch is usually sufficient to saturate the downslope soils, thus resulting in overland flow, and therefore a short lag between melt and stream flow. Second, the melting patches result in high flows continuing after the principal snowmelt period, with characteristic diurnal stream flow peaks. The resulting regime is similar to that found in glacierized basins (Marsh and Woo, 1981). Lewkowicz and French (1982) suggest that the maintenance of the wet conditions downslope of drifts, and the occurrence of overland flow, indicates the applicability of the partial area theory (Betson, 1964) of runoff generation to arctic environments. Overland flow may also dominate in a very different set of circumstances. For example, in subarctic Quebec, Dunne *et al.* (1976) reported that in both forested and treeless areas, where concrete frost is present and the surface is covered by a thin ($<0{\cdot}1$ m) lichen mat, low soil infiltration resulted in lateral flow through a saturated layer at the base of the snowpack with no interaction with the underlying mineral soils.

In contrast to the above conditions, in many areas of the subarctic and arctic tundra, the treeless terrain with a continuous organic cover (NRC, 1988) experiences little or no overland flow, due to a thicker ($\sim 0{\cdot}5$ m) organic layer that includes peat (e.g. Dingman, 1973; Slaughter and Kane, 1979; Quinton, 1997; Carey and Woo, 1997). In this case, subsurface flow normally dominates since the relatively thick organic layer has both a large water storage capacity and a frozen or unfrozen infiltration rate which far exceeds the rate of input from snowmelt or rainfall. In most cases, the water table does not rise above the surface, but resides in the organic layer in areas near the stream throughout the summer (Slaughter and Kane, 1979; Hinzman *et al.*, 1993), while in the upslope areas the water table often falls into the low conductivity mineral soils by midsummer (Quinton, 1997). This spatial variation greatly affects runoff since the upland areas may not be able to rapidly contribute runoff (i.e. storm flow) to the stream channel.

To date, few studies have considered these tundra areas where subsurface flow dominates. The objective of this paper, therefore, is to further examine the runoff mechanics in this environment. This will be accomplished by utilizing field measurements to demonstrate the relative importance of the major processes controlling runoff from the hillslopes adjacent to the stream channel, and from the basin uplands. Based on this information, a conceptual framework that is appropriate for representing the major features of the stream channel hydrograph for this type of environment will be developed and implemented in a simple, physically based approach. The conceptual framework will be evaluated by comparing measured and estimated stream discharge, and recommendations for future research will be provided.

STUDY SITE

The study focused on Siksik Creek (Figure 1), a 95·5 ha subcatchment of Trail Valley Creek (68° 44′N, 133° 28′W), located approximately 55 km north-northeast of Inuvik, NWT on the Mackenzie River Delta, and

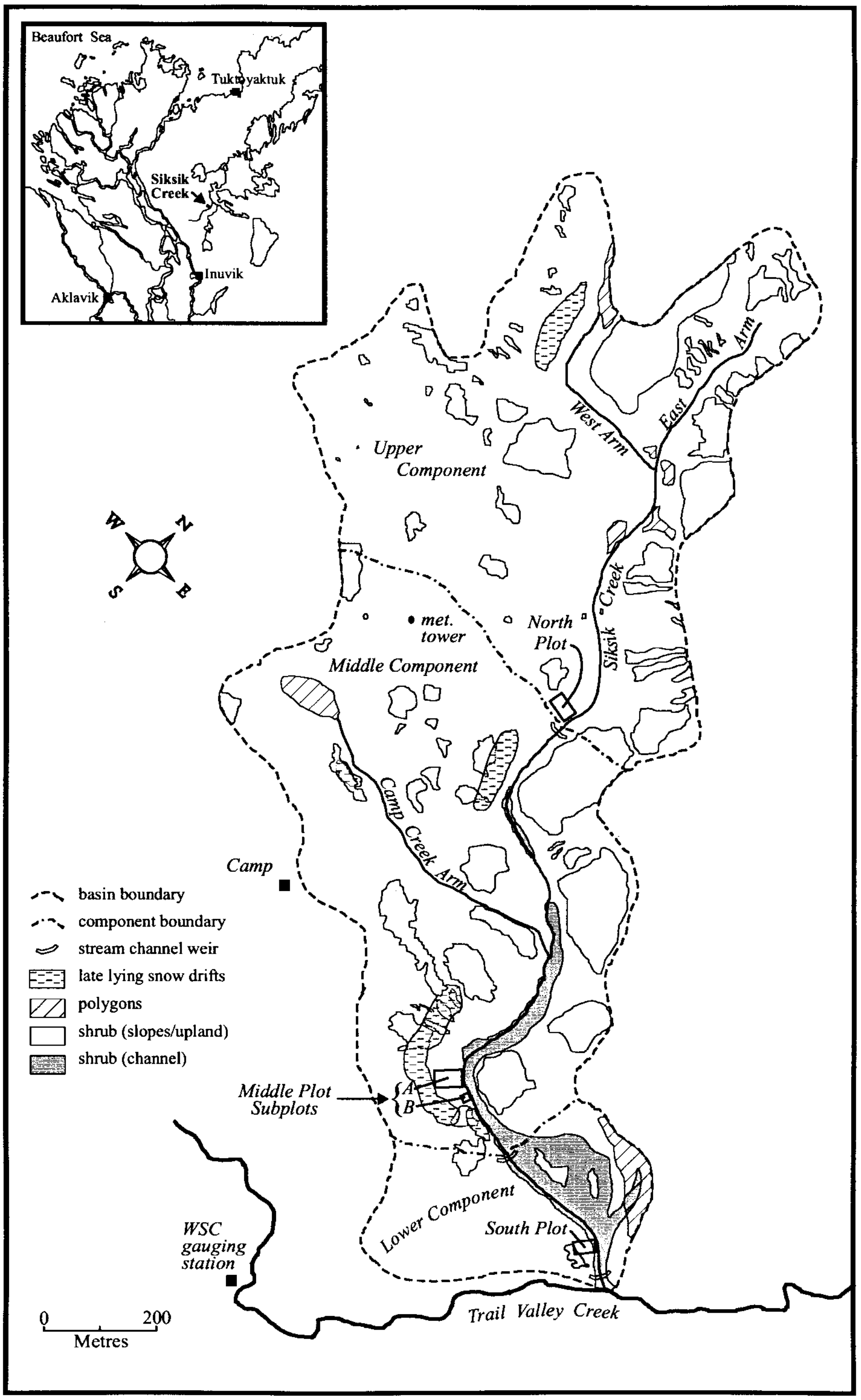

Figure 1. The Siksik Creek basin, showing the location of the hillside study plots, various terrain types, the meteorological tower, and the Water Survey of Canada (WCS) stream gauging station on Trail Valley Creek

80 km south of Tuktoyaktuk, NWT on the Beaufort Sea. The climate in this region is characterized by short, cool summers and long cold winters, with an eight month snow-covered season (Environment Canada, 1982). The mean daily temperature rises above 0 °C in early June, and falls below 0 °C in early October, with a mean annual air temperature of −9·8 °C at Inuvik and −10·9 °C at Tuktoyaktuk. The mean annual precipitation of 266 mm (Inuvik) and 138 mm (Tuktoyaktuk) is comprized of 66% and 47% snowfall respectively. Monthly precipitation is largest in August, September and October, with most precipitation falling as rain in August and September, and snow in October.

Approximately 90% of the annual stream discharge from Siksik Creek occurs during the snowmelt period when the frost table is relatively close to the ground surface (Marsh *et al.*, 1995). After the snowmelt period, the magnitude of hydrological inputs is reduced, the capacity of the soil to store water increases as the active layer thaws, and stream discharge is therefore low. The dominance of snowmelt runoff is typical of arctic regions (Woo, 1986).

The study site is in the zone of continuous permafrost (Heginbottom and Radburn, 1992), and at the northern fringe of the forest–tundra transition zone (Bliss and Matveyeva, 1992), where the tundra vegetation includes low shrubs occurring in isolated patches along the stream channel (Figure 1). The maximum permafrost thickness ranges from 350 to >575 m (Natural Resources Canada, 1995), while maximum observed active layer depths range between 0·4 and 0·8 m depending on aspect and soil type. Because of the extreme cryoturbation associated with periglacial processes, the ground surface is dominated by mineral earth hummocks. In the study area, these hummocks have a diameter of between 0·4 to 1·0 m, and their crests rise between ~0·1 and 0·4 m above the surface of the inter-hummock area. The inter-hummock area is composed of peat ranging in thickness between ~0·2 m and 0·5 m. The upper ~10 cm of the peat is comprized of living vegetation rooted in weakly decomposed plant material. Below this, the peat is composed of moderately to very strongly decomposed plant material with plant structures much less distinct. The inter-hummock area is arranged into a network of interconnected channels that collectively form the hillslope drainage network. Some inter-hummock channels contain soil pipes with diameters of ~5 to 10 cm. These soil pipes are prevalent on hillsides with slopes of 0·08 or higher, but are uncommon on slopes of less than 0·04 (Quinton and Marsh, 1998b). On hillslopes where pipes occur, runoff is characterized by a high degree of interaction between matrix and pipe flow, so that over a distance of a few metres, runoff can alternate between these two flowpath types (Quinton, 1997).

The elevation of Siksik Creek ranges between ~60 m a.s.l., at the confluence with Trail Valley Creek, and ~100 m a.s.l. at the northern edge of the catchment. Over this elevation range, the basin has three distinct components that differ with respect to physiography and surficial geology. The Middle and Upper basin components (Figure 1) are mostly composed of rolling moraines of early Wisconsonian origin with sediment (glacial till, gravel, sand) thicknesses of between 4 and 12 m, and a topography severely effected by thermokarst (Rampton, 1987). Unlike the rest of the basin, the stream channel in the Upper Component is not incised. The Lower basin component is relatively flat, containing 0·5–3·0 m of colluvial clay, silt, sand and rubble deposited during the Holocene by channel, flood basin, overbank, and debris flows (Rampton, 1987).

METHODOLOGY

Hydrology measurements were made at four small (~1000–2000 m^2) experimental hillslope plots: North, Middle-A, Middle-B and South plots (Figure 1) in 1993 and 1994. In the inter-hummock area at each plot, daily measurements of the depth to the water table were made. Manual measurements of water table depth were made daily at observation wells located at various distances from the stream edge. In addition, water table depth was measured continuously with a water level recorder at a single well. A visual survey was conducted daily along the Siksik Creek stream channel in order to determine where surface flow was entering the stream. At the plots, a continuous record of surface runoff was obtained near the stream bank using plywood surface flow collectors inserted to a depth of less than 5 cm. The collectors, which spanned ~20 m, were installed to divert surface flow through a flume box equipped with a 90° v-notch and a stage recorder. Discharge from the flume boxes was measured volumetrically during timed runs. At Middle Plot-A, the only plot with significant surface flow, the collected surface flow did not occur over the entire 20 m width of the flow collectors. Rather, surface flow was concentrated in a single ephemeral 'rill' extending from the lower edge of the snow drift to the stream channel. As a result, the measured flow through the inter-hummock channels was entirely subsurface flow, with no interaction with surface flow. Discharge from the stream bank outlets of several soil pipes was also measured volumetrically during timed turns. Where possible, pipe flow was also measured continuously using a tipping bucket apparatus connected to a Campbell 21X data logger.

Tracer tests were conducted at Middle Plot-A and Middle Plot-B in 1993, and at the North and South plots in 1994 (Figure 1), in order to determine the discharge through inter-hummock channels at different times during the thaw period. Approximately 80 litres of KCl solution (~100 mg L^{-1} Cl^{-1}) was poured onto the surface along a ~10 m line traversing each plot, at a distance of ~10 m upslope of the stream bank. Tracer concentrations were measured in eight, roughly evenly spaced inter-hummock channels at each plot, using chloride sensing micro-electrodes (Farrell *et al.*, 1991) inserted into the peat matrix in the middle of the liquid saturated layer. The sensors were connected to a Campbell Scientific 21X data logger for continuous measurement. Readings were made every 60 seconds and averaged and recorded every 15 minutes. At no time was surface flow observed between the tracer application lines and the stream bank.

Subsurface discharge (Q_C) through the inter-hummock channels was then calculated from

$$Q_C = A_S \eta_A \nu \tag{1}$$

where A_S is the cross sectional area of the saturated layer measured directly at each sensing location, η_A is the active porosity in this layer, and ν is the average velocity through the inter-hummock channel. Since η_A varies with depth, a value representing the elevation of the middle of the saturated layer, was taken from an analyses of changes in η_A with depth (Quinton, 1997). The average velocity ν was calculated from:

$$\nu = \frac{L_X}{t_C} \cdot T_X \tag{2}$$

where L_X is the straight line distance between the tracer application and sensing locations, 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. T_X is the average tortuosity of the inter-hummock channels resulting from the presence of mineral earth hummocks. At Siksik Creek, T_X ranges between 1·1 and 1·5, sand the average value for each plot was taken from Quinton and Marsh (1998a).

In order to quantify variations in peat thicknesses at each plot, ~15 soil pits were excavated at points along a transect perpendicular to the stream bank. Each transect included points at hummocks and in the inter-hummock area. At each soil pit, the total thickness of the peat and its upper and lower layers were recorded, and soil structural features (e.g. macropores or soil pipes) noted. In order to determine how representative the study plots are of the Trail Valley Creek basin, peat thicknesses were measured along nine transects located throughout the Trail Valley basin. These transects extended from the stream bank to an upslope location where the thickness of the peat had stabilized to a minimum value. The depth of the interface between the peat layer and the underlying mineral sediment was easily detected with a graduated steel rod, because of the large difference in the density of the two soil types. The accuracy of the measurements was checked several times on each transect by excavating soil pits for direct observation.

In order to relate plot scale discharge to basin runoff, Siksik Creek discharge was also measured at the downstream end of each basin component (Figure 1). Because the Siksik Creek channel is wide and shallow, and choked with snow early in the season, normal stage measurement techniques were inappropriate, and a reinforced plywood weir, approximately 10 m in width, was installed at each site. Each weir was equipped with a metal-plated 90° v-notch and a stilling well with a stage recorder. Discharge was calculated from the stage above the v-notch using formulae provided by Dingman (1984). To check the accuracy of the weir equations, direct measurements of flow through the v-notch were made regularly using Price (1210 AA and 1205) current metres. Late in the study periods, flow at the notch was low enough to permit volumetric discharge measurement during timed runs by catching all flow.

Standard micro-meteorological measurements (air temperature, relative humidity, wind speed and net radiation) were measured every half hour and recorded by a Campbell 21X data logger at the meteorological station in the Upper Component of the basin (Figure 1). Air temperature was measured with a 76 μm diameter thermocouple, relative humidity was estimated from an aspirated Vaisala HMP35CF capacity sensor, wind speed was measured by an NRG type 40 anemometer field calibrated against a Qualimetrics micro-response 2032 anemometer, which had been calibrated in a wind tunnel, and net radiation from a Middleton CN-2 net radiometer. The late-winter snow water equivalent storage in the basin was computed from a systematic snow depth and density survey of 430 samples that included measurements within all the major terrain types (Marsh *et al*., 1995).

HILLSLOPE PLOT OBSERVATIONS

Soil properties

Figure 2 shows soil profiles at Middle Plot-A with the ground surface set to a common datum. At the time of measurement (23 July, 1993), the frost table was still within the lower peat layer of many of the soil profiles, and as a result, the underlying mineral sediment is not shown in all profiles. The average thickness of the upper peat layer was approximately 10 cm at each plot, but exceeded 20 cm at some profiles (Figure 2). Macropores were observed primarily in the lower peat layer (i.e. the layer of moderately to strongly decomposed peat). At Middle Plot-A and B, many of these macropores were elongated in the horizontal direction, forming soil pipes.

The bulk saturated hydraulic conductivity of inter-hummock channels computed for the ~10 m distance over which the subsurface velocity was measured during the tracer tests, ranged between ~22 and 1810 m d^{-1} (average ~448 m d^{-1}, N = 41) when the water table was in the upper peat layer, and between ~0·7 and 212 m d^{-1} (average ~27 m d^{-1}, N = 30) when the water table was in the lower layer (Quinton, 1997). In the upper layer, hydraulic conductivities were typically <1000 m d^{-1}, with only one value (1810 m d^{-1}) considerably higher. All the computed hydraulic conductivities fall within the range of values for peat reported by other studies (e.g. Ivanov, 1975; Hoag and Price, 1997). The hydraulic conductivity of the

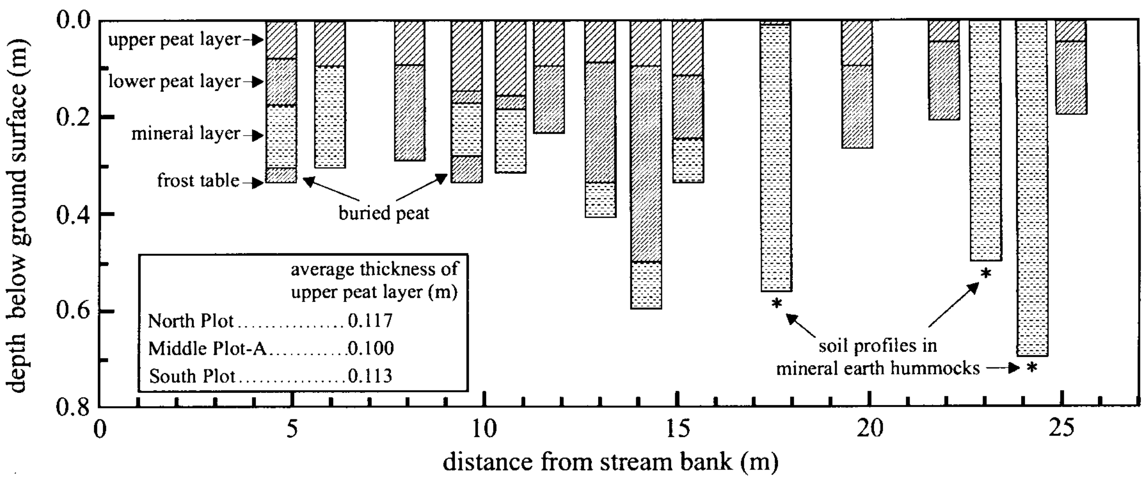

Figure 2. The major layers of soil between the ground surface and the frost table measured on 25 July, 1993 at different distances from the stream bank at Middle Plot-A. All but three profiles were in the inter-hummock area. The average peat thickness of the upper peat layer at this plot, and at the North and South plots is tabulated

underlying, unfrozen, mineral sediment ($\sim 10^{-2}$ to 10^{-3} m d^{-1}) is two to three orders of magnitude lower than that of the basal peat (Quinton and Marsh, 1998b).

Runoff observations

As the active layer thaws and the input of meltwater reduces, the position of the saturated layer changes, as does the relative importance of the various hillslope runoff pathways. Other than a brief (~two day) surface runoff event at the North Plot in 1994, Middle Plot-A was the only plot over the two year study period, where surface runoff was observed. At this site, in both years, surface flow did not commence until most of the ground surface had become free of snow (Figure 3), and the remaining snow, the deep, late-lying drifts, began contributing meltwater. The surface runoff at Middle Plot-A occurred exclusively along the ephemeral rill at a location separate from where the tracer tests and pipe flow measurements were conducted.

Approximately 94% of the cumulative annual surface flow from Middle Plot-A in 1993 and 90% in 1994 was generated by snowmelt, with the majority originating from the snow drift (Figure 1), and with the remainder resulting from late-summer rain events. Prior to the disappearance of the snow drift from the Middle Plot, which occurred on approximately 18 June in both 1993 and 1994, the cumulative rainfall was only ~9 mm in 1993 (five events), and ~10 mm (six events) in 1994. As a result, rainfall played a minor role in generating runoff during this period. From the daily surveys along the Siksik Creek stream channel, no surface flow was observed entering the stream except for downslope of the snow drift at Middle Plot-A. Surface runoff from this plot (as shown in Figures 3 and 4), accounts for approximately 0·7% of the Siksik Creek stream discharge during the snow melt runoff periods of 1993 and 1994. The snow drift generating this surface flow had the largest surface area among the drifts in the basin, and occupied ~1·1% of the basin area. If surface flow was generated during melt at the same rate per unit area at all drifts, an area representing ~2·4% of the Siksik Creek basin (Figure 1), the estimated contribution to the cumulative stream discharge of the melt period would be only ~1·6%. This, along with the daily visual surveys that found no other surface runoff entering the stream channel, suggests that the basin-wide contribution of surface runoff to stream flow is very small.

Despite the small contribution of surface flow, observations show that hillslopes can deliver subsurface runoff to the stream channel relatively quickly. For example, at Middle Plot-A, daily fluctuations in the water level in the inter-hummock channels, and flow measured directly from a pipe outlet, suggests that water drains below the surface in daily waves generated by melt of the late-lying snow drift (Figure 3). Comparison of the time of the daily peak of surface flow to that of the daily peak water table elevation (Figure 3a) and to the time of peak pipe discharge (Figure 3b), suggest that when the water table is relatively close to the ground surface, the flow velocity through the peat matrix is approximately as high as for pipe flow and surface flow.

Results from the tracer experiments show that as the elevation of the saturated layer descends through the active layer, the average velocity of flow through the saturated layer reduces by two to three orders of magnitude (Table I). As a result, the discharge through inter-hummock channels decreases abruptly as the elevation of the saturated layer declines. For the tracer tests prior to 18 June, 1993, the large moisture supply

Table I. The average of the velocities computed from equation (2) at each plot for different depths (d) from the ground surface to the middle of the saturated peat layer. *Although the depth to the middle of the saturated layer had slightly increased, the water table was closest to the surface when this estimate was made

North		Middle-A		Middle-B		South	
d (m)	v (m d^{-1})	d (m)	v (m d^{-1})	d (m)	v (m d^{-1})	d (m)	v (m d^{-1})
0·077	28·6	0·097	50·3	0·097	92·9	0·09	34·7
0·153	9·2	0·099	27·7	0·099	153·4*	0·115	5
0·168	2	0·119	13·6	0·119	2·3	—	—
—	—	0·205	1·3	0·205	1·1	—	—
—	—	0·322	0·6	0·32	0·5	—	—

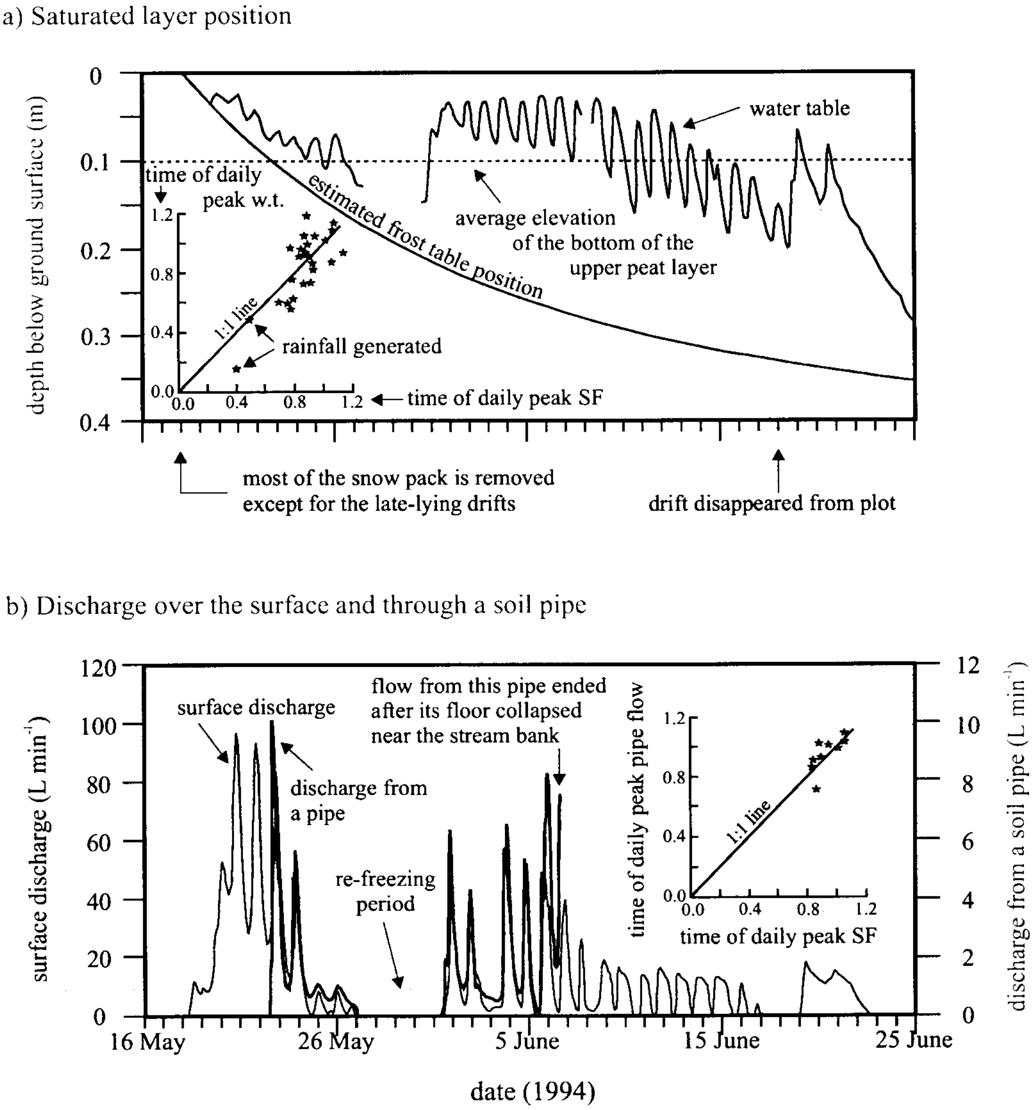

Figure 3. The position of (a) the saturated layer in the peat of the inter-hummock area and (b) the discharge hydrographs of surface flow and of flow from a pipe outlet at Middle Plot-A in 1994. Note that the units for the two inset graphs are days. A value between 1·0 and 1·2 implies that the peak occurred between midnight and 5:00 AM. SF refers to surface flow in both inset figures

due to the presence of the melting snow drift upslope of the plot, and the closeness of the saturated layer to the ground surface, enabled a relatively high level of discharge through both types of inter-hummock channels — those with soil pipes, and those without (Figure 4). During this period of high subsurface discharge, water drained from the plot in diurnal waves, as shown in Figure 3, and as a result, the stream hydrograph fluctuated diurnally (Figure 4).

Shortly following 18 June, 1993, the snow drift upslope of the plot had entirely melted, and as a result, surface flow in the single rill ended, and the water table subsided below the highly conductive upper peat layer. By the tracer experiment of 5 July, the water table occupied the lower peat layer, but was still above the elevation of most soil pipes. Therefore discharge through the inter-hummock channels containing these

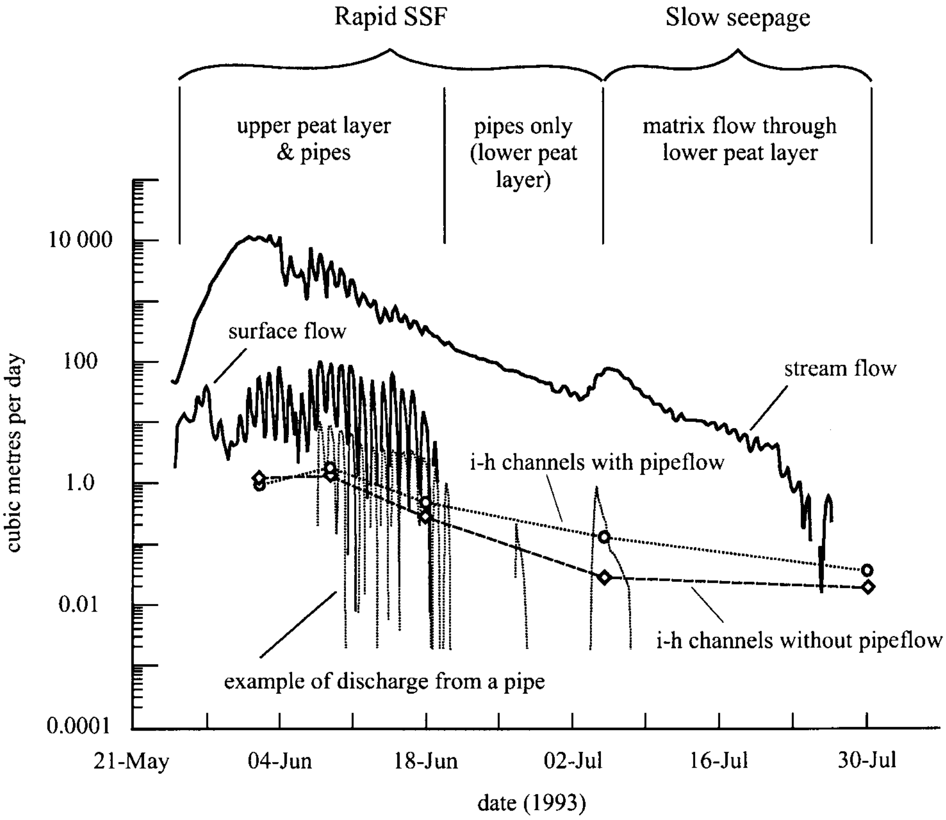

Figure 4. The average subsurface discharge through inter-hummock (i–h) channels with pipes (circles) and without pipes (diamonds) estimated from the five tracer experiments in 1993 (2, 9, 18 June; 5, 30 July). In order to increase the number of observations of subsurface discharge through both types of inter-hummock channel, the data from Middle Plot-A and B were combined. Also shown are Siksik Creek stream discharge at the downstream end of the Middle Component (Figure 1), surface runoff from Middle Plot-A, and an example of flow from a single pipe

features continued at a relatively high rate (Figure 4). Between 5 and 30 July, rapid subsurface flow through the soil pipes ended, since drainage and active layer development displaced the water table below the elevation of most pipes. As a result, the average discharge through both types of inter-hummock channel converged to approximately the same low rate (Figure 4). Presumably this level of inter-hummock channel discharge to the stream channel was insufficient to maintain stream flow, since by the final tracer experiment at this plot (30 July, 1993), discharge from Siksik Creek had stopped.

Darcian flow

Although subsurface flow through the peat matrix and soil pipes can be rapid, calculations of the Reynolds number (N_R), shows that N_R is always well below 10, the upper limit of laminar flow through porous media. For example, N_R calculated for the lowest (0·16 m d^{-1}) and highest (353·5 m d^{-1}) velocity observed in this study, using the geometric mean pore diameters of the peat matrix for the position of the saturated layer at the time of these observations (0·22 mm for the lowest velocity, and 0·46 mm for the highest (Quinton, 1997)), shows that N_R ranges between $3{\cdot}14 \times 10^{-4}$ and 1·45. This suggests that Darcy's law is applicable in describing subsurface flow at the study plots.

HILLSLOPE RUNOFF COMPUTATIONS

Computation procedure

Since the above observations show that the majority of stormflow and total hillslope runoff occurs as subsurface flow through the inter-hummock channels, a physically realistic method of estimating stream flow requires the calculation of inter-hummock channel flow for the range of conditions found in the basin. As an initial test of this approach, we calculated the average inter-hummock channel discharge weighted by the total number of inter-hummock channels. The objective of this approach is not to accurately determine stream flow, but rather to determine if such a procedure explains the primary features of stream flow in this environment. As a first approximation, the hillslope characteristics of Middle Plot-A were used for this computation, and the following summarizes the procedures used.

The thickness of the saturated layer within the inter-hummock peat (S_L) was calculated from

$$S_L = W_T - F_T \quad \text{if} \quad F_T > P_T$$

or

$$S_L = W_T - P_T \quad \text{if} \quad F_T < P_T \tag{3}$$

where W_T is the water table elevation, F_T the frost table elevation and P_T is the elevation at the base of the peat. Water level elevation was continuously measured at a single well in the inter-hummock area of Middle Plot-A and F_T was computed as a function of time following the disappearance of the snow cover using the function defined by Quinton (1997). Owing to the highly inter-connected nature of the inter-hummock channels (Quinton and Marsh, 1998a), the water table data represents the inter-hummock channels across the lower section of the hillslope plot. The average subsurface discharge of the inter-hummock channels (Q_C), was then computed on 15-minute time steps from

$$Q_C = K \cdot \frac{\Delta H}{\Delta x} \cdot A_S. \tag{4}$$

For each time step, the hydraulic conductivity (K) was computed empirically as a function of the water table elevation (Quinton and Marsh, 1998a; Quinton, 1997). If the frost table was above the bottom of the peat, the hydraulic conductivity was computed for the position mid-way between the frost and water tables, otherwise it was computed for the position mid-way between the water table and the base of the peat. For the hydraulic gradient ($\Delta H/\Delta x$), a constant value of 0·08, equal to the average slope of the ground surface at Middle Plot-A, was used. This value was found to closely approximate the hydraulic gradient throughout the study period (Quinton, 1997). The cross-sectional area of the organic portion of the saturated layer (A_S) was calculated from

$$A_S = S_L \cdot IHC_W \tag{5}$$

where IHC_W is the average width of the inter-hummock channels (0·45 m). If the water table is in the underlying mineral soil, then $Q_C = 0$.

The total inter-hummock channel subsurface input to the stream channel Q_{SS} was then computed from

$$Q_{SS} = Q_C \cdot IHC_F \cdot (2 \cdot C_L) \tag{6}$$

where IHC_F is the frequency of inter-hummock channels per length of stream bank (0·7 m^{-1}) and C_L is the total length of Siksik Creek (2485 m), which was multiplied by two to account for drainage from both sides of the stream.

Since ~10 to 15% of the total basin snowpack occurs in the Siksik Creek stream channel (Marsh *et al.*, 1995; Marsh and Pomeroy, 1996), it is necessary to include the meltwater input from this snow in an estimate of stream flow, since it is not included in flow from the hillslopes (i.e. Q_{SS}). The melt water from the stream channel (C_M) was estimated from

$$C_M = M \cdot C_L \cdot C_W \tag{7}$$

where M, the average daily snow melt rate, was estimated from the bulk aerodynamic method (Marsh and Pomeroy, 1996), and C_W is the average width of the snowpack occupying the stream channel (20·0 m). This contribution was assumed to continue until the cumulative input from this source equalled the initial snow water equivalent of the snowpack in the stream (350 mm of snow water equivalent). Stream flow was initiated once hillslope runoff (Q_{SS}) began. Once this occurred, stream flow (Q_{STR}) was then estimated from

$$Q_{STR} = Q_{SS} + C_M + C_S \tag{8}$$

where C_S is the liquid storage in the stream channel that accumulated from inputs of melt water from the channel snowpack prior to the input of Q_{SS} to the stream channel. Water was released from this storage term at a constant rate equal to the average rate that C_M entered this storage. From this assumption, the C_S term was depleted by the third day of stream flow. Over this three-day period, C_S accounted for ~25% of the cumulative stream discharge.

Computed discharge

The procedure for estimating stream flow described above captures the primary features of the Siksik Creek hydrograph (Figure 5) for the spring and early summer period of 1994. The date flow began, the peak discharge, the low flow during the re-freezing period and the diurnal variations in flow are similar for both the observed and computed hydrographs. In addition, the cumulative observed and computed discharges for the 40 day period shown in Figure 5 differ by only 8%. These similarities suggest that the computational

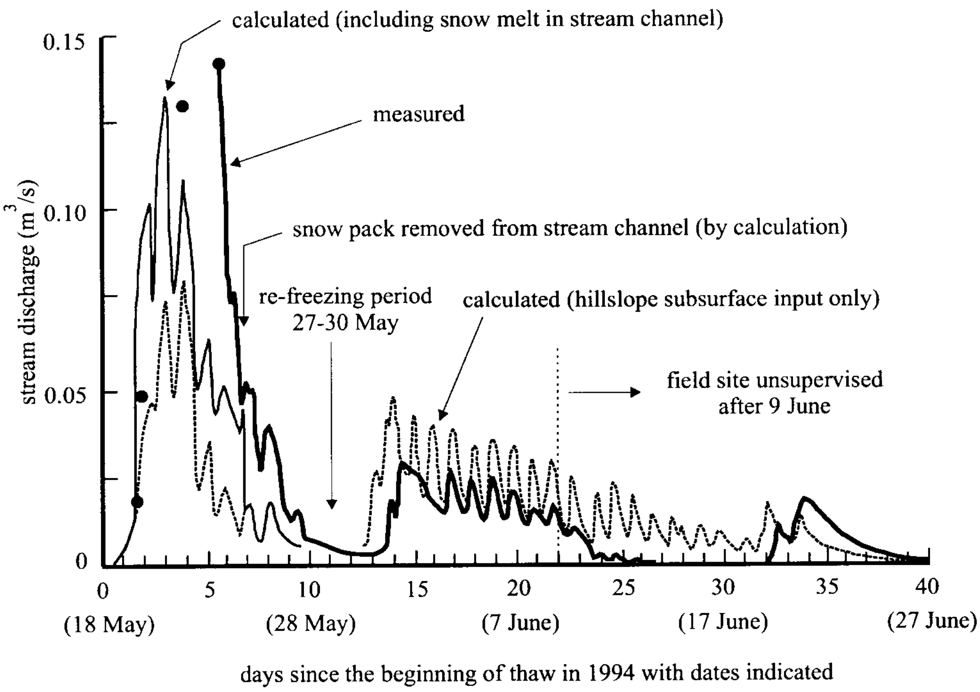

Figure 5. Stream discharge measured at Siksik Creek plotted with stream the computed hydrographs. Approximately seven days after the initiation of stream flow, the channel snowpack was melted, and only a single computed hydrograph is shown

procedure could be a useful conceptual framework for the development of a runoff model for this environment. However, there are numerous features that are not properly described by the computed hydrograph, suggesting that improvements to this procedure are required.

The earlier rise and peak of the computed hydrograph before and immediately after the re-freezing period is likely the result of the stream channel storage not being adequately represented in the computations, and the channel routing not being included at all. As a result, the estimated discharge hydrograph does not account for the lag resulting from the time required for hydrological inputs ($Q_{SS} + C_M$) to reach the basin outlet. Prior to day 10, the calculated contribution from the hillslope inter-hummock channels (Q_{SS}) to Siksik Creek is much smaller than the observed discharge (Figure 5). However, when the meltwater from the deep snowpack in the stream channel (C_M) is added to Q_{SS}, the peak of the computed hydrograph increases by ~50%, and approaches the observed peak discharge. In spite of this, the total flow computed for this period was still approximately 26% less than the observed. This is likely due to no account being made of hillslope runoff entering the stream channel while the hillslopes were still snow-covered. Quinton and Marsh (1998b) found that this process makes a contribution to storage in the stream channel prior to the initiation of stream flow.

Following the re-freezing period of 27 to 30 May (Figure 5), the estimated stream flow peaks were higher than the observed, while the daily minimums were similar, resulting in the total flow being overestimated by ~76%. This was largely due to the effect of the late-lying snow drift upslope of Middle Plot-A, the plot used to drive the computation of stream flow. This drift occupied ~50% of the plot, while the percentage of the basin occupied by such drifts was only 2·4% (Figure 1). Estimating the basin-wide hillslope runoff contribution to the stream channel from the observed rate of drainage from the Middle Plot therefore contributed to the overestimated total stream flow of this period. The agreement between the computed and observed hydrographs was especially poor following day 22 (9 June), due to measurement errors as the field site was left unattended after this date.

While obvious problems exist with the present computations, the basic framework identifies the important computations to be made, and indicates where improvements are required before the framework is used in a distributed hydrological model. For example, a complete model would need to generate the water table elevation and the saturated layer thickness (rather than rely upon measured water table elevations and empirically-defined frost table depths), and would need to do this for a range of typical slope types to account for the spatial variation of this factor. An improved computation of the channel storage term (C_S), which would include hillslope runoff contributions to channel storage, and a channel routing procedure would also be required.

HYDROLOGICAL LINKAGES AMONG BASIN SUBSYSTEMS

The previous sections focused upon inter-hummock subsurface flow as observed at hillslope plots, and channel melt, and demonstrated how the contribution from these sources can be computed and used to estimate stream flow. However, in order to fully model hillslope runoff and streamflow, a better understanding of the role of the runoff processes discussed in the previous sections in determining the basin-wide runoff response is needed. This includes an understanding of the hydrological linkages among the major basin subsystems, namely the stream channel, the near-stream area and the basin uplands. The remainder of this paper uses the understanding of runoff processes gained from the previous sections, to explain the runoff response at the basin scale.

The saturated peat layer

Give the importance of subsurface flow through the peat at Siksik Creek, an important variable influencing stream flow is the areal extent of the saturated peat zone. The proportion of this area connected to the stream channel as a continuous layer of saturated peat defines the source area of total subsurface drainage to the stream, including rapid flow (through the upper peat layer and soil pipes), and slow seepage through the peat matrix of the lower peat layer. One of the factors affecting the areal extent of saturated peat

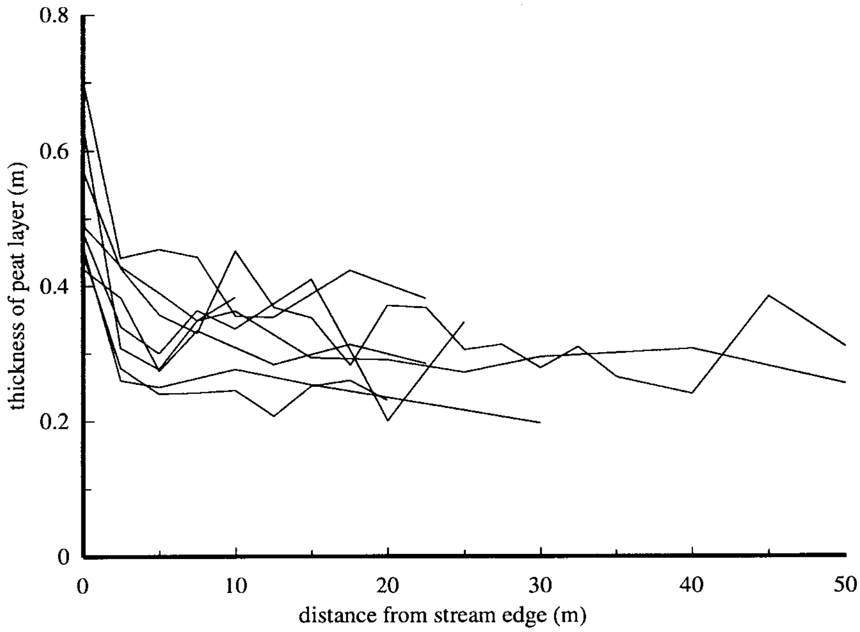

Figure 6. The variation in the thickness of the peat layer with distance from stream channels at transects within the Trail Valley Creek watershed

is the spatial variation of peat thickness, since the frost table, and therefore the water table, remains in the peat for a longer time in areas where the peat is relatively deep. Observations of peat thickness show that it decreases from between ~0·4 to 0·7 m near the stream channel and levels off to ~0·3 m at a distance of ~10 to 40 m upslope of the stream edge (Figure 6). This suggests that hillslopes may be divided into two zones based on the distribution of peat thickness: a zone where the peat is relatively thick, but decreases in thickness away from the channel; and an area further away from the channel where the peat layer is relatively thin. In this paper, we will refer to these two areas as the near-stream area and the upland area. This shallow peat of the uplands extends downslope from the topographic uplands to the narrow near-stream area. The near-stream area occupies ~10% of the total Siksik Creek basin area based on the product of its average width (~40 m) estimated from aerial photographs, and the total stream channel length (2485 m), with the uplands covering the remaining ~90% of the basin (Quinton and Marsh, 1998b).

When the water table in the inter-hummock channels is close to the surface (Figure 7a), as it is during the spring melt period when the thaw layer is thin or after a large summer rain storm, saturated peat extends as a continuous layer throughout much of the basin. During summer dry periods, however, the water table declines, and as a result, the zone of saturated peat is limited to an area close to the stream (Figure 7b). This depiction is based on observations that during the summer the water table within the peat had become discontinuous in the basin uplands with the water table in the peat in relatively flat areas, while in areas of greater inclination, including the slopes leading to the near-stream area, the saturated layer had drained from the peat (Figure 8). As a result, much of the basin uplands is hydrologically disconnected from the stream, given the very low hydraulic conductivity of the underlying mineral sediment. Such variations in the lateral extent of the saturated peat layer have considerable impact on stream flow, and must be considered when modelling runoff in this environment.

The role of rapid subsurface flow in linking basin subsystems

In addition to the lateral extent of the saturated peat layer, stream flow is also influenced by the elevation of the water table, since it controls the subsurface velocity and therefore the time lag time between input

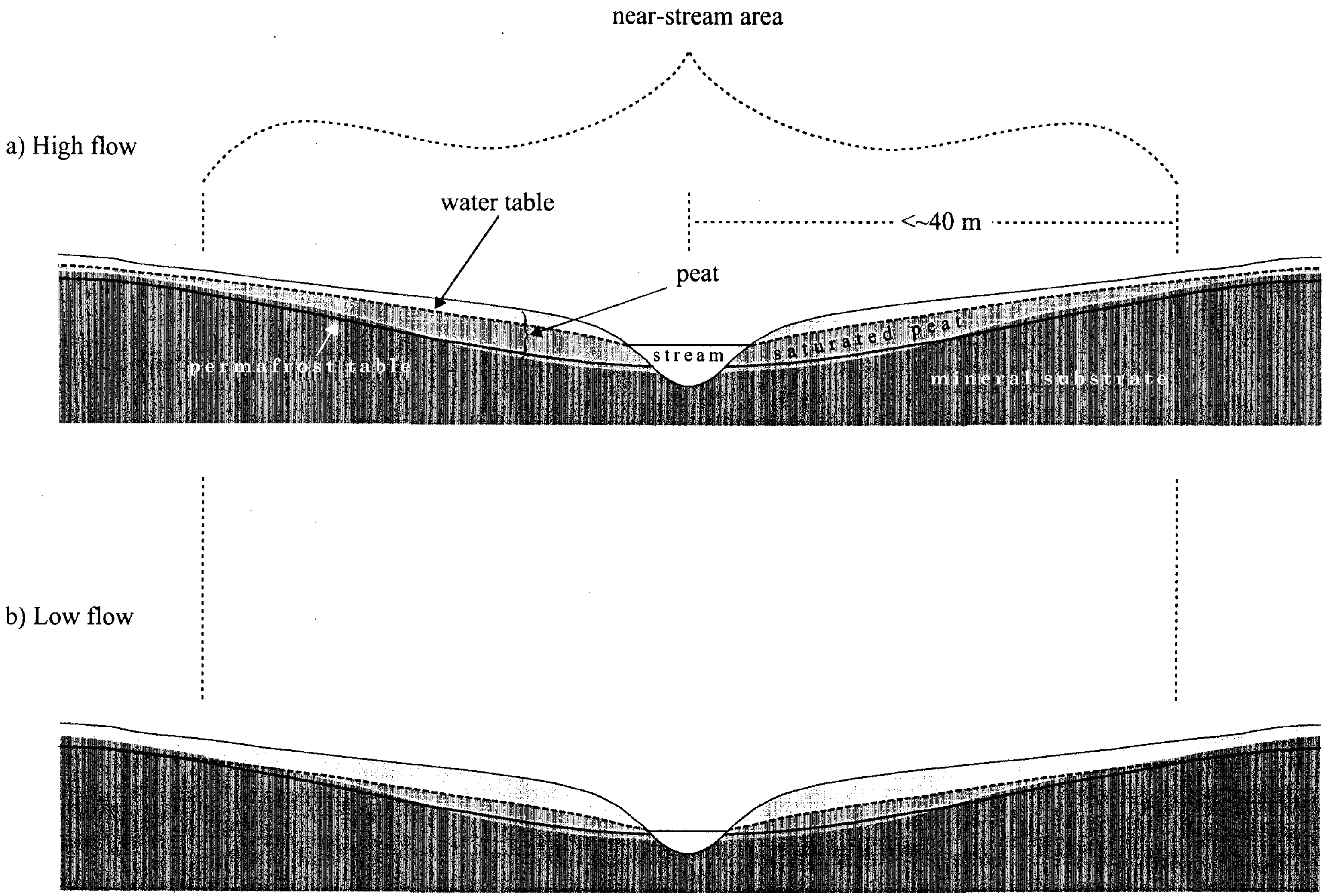

Figure 7. The near-stream area shown schematically in cross-section. The thickness of the peat and of the saturated layer within the peat are greater in the near-stream area than in the uplands. The approximate location of the permafrost table is also shown

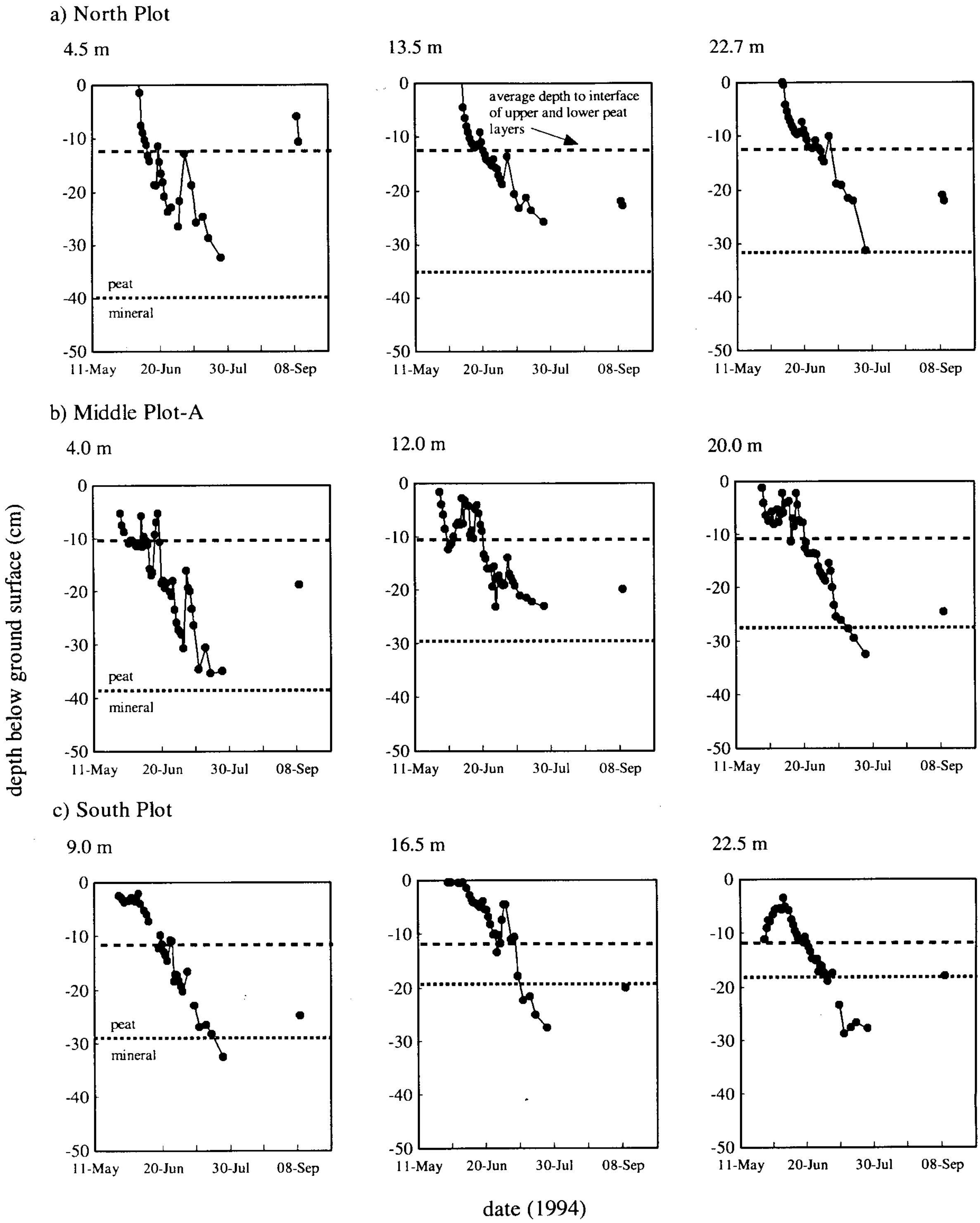

Figure 8. The peat thickness and water table position for three distances from the stream edge at (a) the North, (b) Middle-A and (c) South plots

Table II. The impact of the near-stream area on stream flow. NSA = near-stream area; UPL = upper peat layer; w.t. = water table; SSF = subsurface flow; Q = stream discharge.

Flow Regime	Water table position		NSA-Upslope SSF Connection	NSA ability to convey Rapid SSF	Source area for basin Q
	NSA	Upland			
High	Upper peat layer	Upper peat layer	Rapid SSF ↙ ↘ UPL Pipes	High	NSA & Upslope (i.e. large)
Low	Lower peat layer	Lower peat layer (or w.t. in mineral Substrate)	Slow Seepage	Low	Within NSA (i.e. small)

(from rain or snowmelt) and output to the stream channel. High subsurface flow rates also enable subsurface stormflow to reach the stream channel from greater distances upslope. In the near-stream area, the elevation of the water table determines the type of linkage that this area provides between the stream channel and the uplands. For example, when the water table is in the highly conductive upper peat layer (Table II), water flows into and through the near-stream area rapidly, and therefore this area offers little attenuation to runoff. However, when the water table is in the lower peat layer, water draining into the near-stream area flows toward the stream as slow seepage. Under this condition, if the drainage from the upland into the near-stream area were to increase (e.g. in response to an input event), the near-stream area can delay the lateral expansion of the source area over which storm flow is generated, since the relatively large water storage capacity of this area must be sufficiently filled before the source area can expand. It is emphasized that stormflow production does not require that the storage capacity be entirely filled (i.e. as occurs when the water table is at the ground surface) as envisioned in the variable source area concept (Betson, 1964; Hewlett and Hibbert, 1967) since a water table rise to an elevation where rapid subsurface flow can occur (i.e. through soil pipes, or the upper peat layer) is sufficient to produce surface flow-like velocities (Figure 3).

Spatial variation of the near-stream area width

The width of the near-stream area influences hillslope drainage, since as the width increases, so does the water storage capacity. For example, if the hydrological response of two channel sections are compared, the high flow regime would be initiated earlier, and would last longer at the section where the near-stream area is smaller, if other factors are equal. This suggests that the variability of the near-stream area width should be accounted for when attempting to model runoff from basins like Siksik Creek. Some sense of the degree of variability of the width of the near-stream area can be taken from this study. Observations show that it varies from ~10 to 40 m, while at the study plots represented in Figure 8, the width of the near-stream area is greatest at the North Plot, where the peat is relatively thick, and the water table remains within the peat over the greatest distance from the stream edge. The near-stream area extends further upslope at Middle Plot-A (between 12 and 20 m from the stream edge) than at the South Plot where the water table eventually falls below the peat at a distance of only 9 m from the stream edge. Observations at Siksik Creek suggest that the near-stream area is broader in areas where the inclination and length of the hillslopes are relatively low (e.g. North Plot).

Spatial variation in the duration of high subsurface flow rates

In addition to spatial variables in the factors discussed above, variations in hillslope attributes, including their length and inclination, the water equivalent of the snowpack, and the rate of snowmelt also contribute to variations in the flow regimes of different hillslopes. Figure 9 shows the duration of the high flow regime ($\sim$0·1–1·0 $m^3\ d^{-1}$) at each plot, as indicated by the length of time that the water table remains in the highly

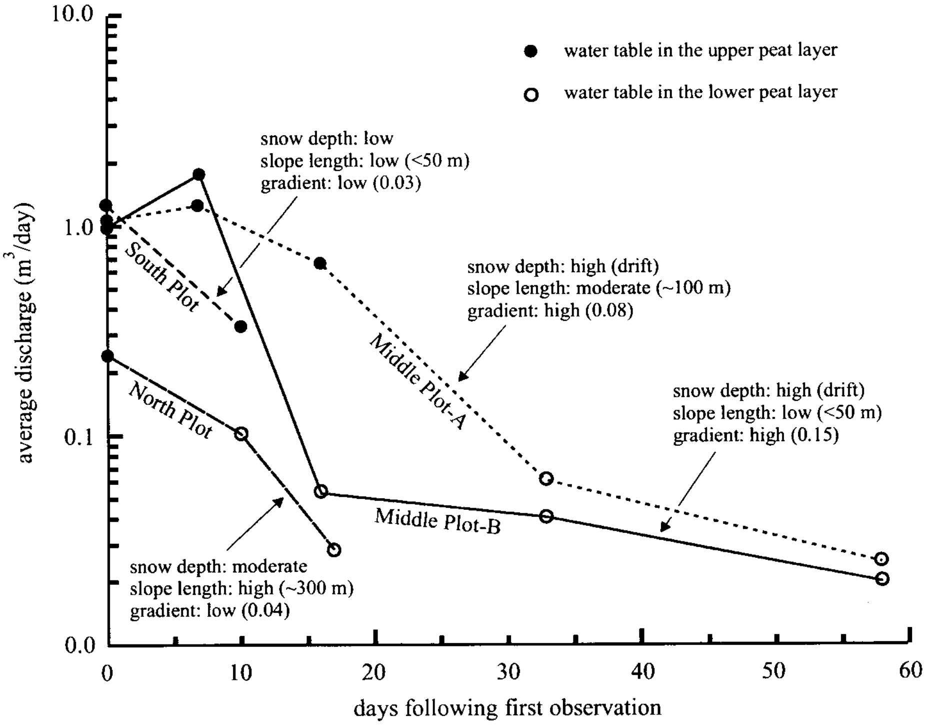

Figure 9. Temporal variations in the average subsurface discharge from inter-hummock channels from the hillslope plots at Siksik Creek. On the horizontal axis, time is expressed in days following the first observation rather than dates, since the tracer tests were conducted at Middle Plot-A and B in 1993, and at the North and South plots in 1994

conductive upper peat layer. Figure 9 also identifies physical attributes of each hillslope which help to explain the variation in the duration of the high flow regime among the plots. At Middle Plot-A and B, the presence of a melting drift upslope kept the water table within the upper peat layer for an extended period. Prior to day 10, subsurface runoff velocities were higher at Middle Plot-B than A owing to the higher slope angle at the former (Figure 9). However, the late-lying snow drift located upslope of Middle Plot-B melted soon after day 10, and as a result the water table subsided into the lower peat layer, and the hillside adjusted to a low flow regime ($<0{\cdot}1$ m^3 d^{-1}).

Discharge from the North and South plots adjusted more rapidly to a lower level, since these plots had relatively low slope angles, and were not affected by deep, late-lying snow. The initial decline to the low flow regime is more gradual at the North Plot, due to the greater slope length (and therefore larger drainage area), and the deeper snow at this plot. In cross-section, the slope of the North Plot is relatively constant over its length. However, the South Plot is convex in cross-section, with the slope decreasing with distance between the stream edge (where the gradient is similar to that of Middle Plot-B), and ~50 m upslope, beyond which the ground surface is flat, or has very little gradient. Therefore once the runoff water supplied by the shallow snowpack of the South Plot drained from the upper peat layer, it is likely that lateral drainage from the large flat area occupying most of the Lower Component (Figure 1) was greatly reduced, and that hillslope drainage became largely restricted to the areas of higher slope close to the stream.

CONCLUSIONS

It was found that subsurface flow through the peat layers of the inter-hummock channels is the dominant mechanism of runoff to the stream, and that subsurface flow through the highly conductive upper peat layer and soil pipes can be as rapid as surface flow. As the elevation of the saturated layer declines over the thaw period, the average flow velocity through it decreases by two to three orders of magnitude. As a result, discharge through the inter-hummock channels decreases abruptly as the elevation of the saturated layer lowers. Rapid subsurface flow can occur through the matrix of the upper peat layer and through soil pipes. When the water table is close to the ground surface, discharge is relatively high through inter-hummock channels with and without soil pipes. When the water tale subsides into the lower peat layer, the discharge rate declines to a lower level, except for inter-hummock channels with soil pipes, where discharge remains relatively high for an extended period.

Based on the observations at the hillslope plots, a procedure for computing stream discharge was introduced, and then used to compute stream discharge for the spring and early summer period of 1994. The computational procedure was shown to capture the primary features of the measured Siksik Creek hydrograph. The similarity between the computed and measured discharge suggests that the computational procedure would be a useful conceptual framework for a distributed hydrological model for this environment. However, such a model would also need to account for additional factors operating at spatial scales larger than the hillslope plots. These include the areal extent of the saturated peat layer in the basin, the elevation of the water table in the near-stream area, and the spatial variation of the width of the near-stream area. These factors influence the hydrological linkages among the three major subsystems of the basin (the stream channel, the near-stream area and the uplands), and therefore the rate and variability of drainage to the stream channel. For example, when the water table is close to the surface, not only is flow rapid (because of the high permeability of the upper peat layer), but the source area for the production of storm flow is relatively large, since a high water table position in the near-stream area enables the source area for subsurface stormflow to expand outward into the basin uplands. However, when the water table is in the lower peat layer, the near-stream area would delay a hydrograph response in the stream channel, since water entering the near-stream area could only flow to the stream as slow seepage through the matrix of the lower peat layer. This condition would continue until the water table rises to a level where rapid subsurface flow can occur (i.e. soil pipes or the upper peat layer).

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