Evaluating snowmelt runoff generation in a discontinuous permafrost catchment using stable isotope, hydrochemical and hydrometric data*

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Abstract Research on snowmelt runoff generation in discontinuous permafrost subarctic catchments has highlighted the role of: (i) permafrost in restricting deep percolation and sustaining near-surface water tables and (ii) the surface organic layer in rapidly conveying water to the stream. Conceptual models of runoff generation have largely been derived from hydrometric data, with isotope and hydrochemical data having only limited application in delineating sources and pathways of water. In a small subarctic alpine catchment within the Wolf Creek Research Basin, Yukon, Canada, snowmelt runoff generation processes were studied during 2002 using a mixed methods approach. Snowmelt timing varied between basin slopes, with south-facing exposures melting prior to permafrost-underlain north-facing slopes. The streamflow freshet period begain after 90% of snow had melted on the south-facing slope and coincided with the main melt period on the north-facing slope, indicating that contributing areas were largely defined by permafrost distribution. Stable isotope ($\delta^{18}O$) and hydrochemical parameters (dissolved organic carbon, specific conductivity, pH) suggest that, at the beginning of the melt period, meltwater infiltrates soil pores and resides in temporary storage. As melt progresses and bare ground appears, thawing of soils and continued meltwater delivery to the slopes allows rapid drainage of this meltwater through surface organic layers. As melt continues, soil thawing progresses and pre-event water mixes with melt water to impart streamflow with a gradually decreasing meltwater contribution. By the end of the melt period, the majority of water reaching the stream is displaced water that has resided in the catchment prior to melt. For the entire study period, approximately 21% of freshet was supplied by the snowpack, and the remaining majority was pre-melt water stored in the catchment slopes over-winter and displaced during melt. Hydrochemical data support hydrometric observations indicating the dominant flow pathway linking the slopes and the stream is through the organic horizon on permafrost-underlain slopes.
Keywords Discontinuous permafrost; runoff; stable isotope; snowmelt; subarctic

Introduction

In north-western North America, hydrologic studies in the zone of continuous and discontinuous permafrost have yielded conceptual models of streamflow generation and revealed the flow pathways whereby water moves from the hillslopes to the drainage network (Dingman 1971; Santeford 1979; Kane *et al.* 1981; Slaughter *et al.* 1983; Hinzman *et al.* 1993; McNamara *et al.* 1997; Quinton and Marsh 1999; Carey and Woo 2001a). Several factors that distinguish the runoff hydrology of permafrost regions from more temperate regions have been identified as: (1) snowmelt is the major hydrological event, often conveying up to half of the annual precipitation inputs to the catchment in a two to four week period (McNamara *et al.* 1998; Carey and Woo 2001b), (2) deep drainage is restricted where permafrost is present, enhancing near-surface water tables (Slaughter and Kane 1979; Kane

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et al. 1989; Quinton and Marsh 1999; Carey and Woo 2001a), (3) the widespread occurrence of capping organic soils allow the rapid translocation of water to the stream when the phreatic surface resides in this upper transmissive layer (Quinton *et al.* 2000; Carey and Woo 2001a), and (4) matrix bypass mechanisms such as soil pipes (Gibson *et al.* 1993; Carey and Woo 2000) and inter-hummock flow (Quinton and Marsh 1998) may convey significant amounts of water during the melt period and wet antecedent conditions. The significant majority of these factors have been identified from hydrometric observations, as runoff generation research using isotopic and hydrochemical methods are scarcer and have not been closely tied with the development of conceptual runoff models (e.g. Obradovic and Sklash 1986; Cooper *et al.* 1991, 1993; Gibson *et al.* 1993; McNamara *et al.* 1997; Carey and Quinton, 2004).

The reliance on hydrometric methods in permafrost systems contrasts work in more temperate areas where over the past several decades methods employing isotopic and hydrochemical data have been used together with hydrometric information to conceptualize runoff processes through identification of sources, residence times and pathways of water as it moves from the catchment to the stream (e.g. Maulé and Stein 1990; Maloszewski and Zuber 1992; Bonell 1993; Buttle 1994; Brown *et al.* 1999). Typically, stable isotopic ratios are used to define the sources of water and dissolved solutes provide information on flow pathways (e.g. Hooper and Shoemaker 1986; Maulé and Stein 1990; Rice and Hornberger 1998; Hoeg *et al.* 2000; Ladouche *et al.* 2001). Using this integrated information, conceptual models of runoff generation have been developed for a wide variety of global environments (e.g. Bonnell *et al.* 1998; Gibson *et al.* 2000; McGlynn *et al.* 2002).

In permafrost catchments, there is little debate that pre-event water stored within the basin dominates summer stormflow hydrographs generated by rainfall (McNamara *et al.* 1997; Carey and Quinton 2004), but there remains significant uncertainty and conflicting evidence as to whether this pre-event water is the principal contributor to the freshet hydrograph when large volumes of meltwater are released and streamflow rises to its maximum annual flows (Obradovic and Sklash 1986; Cooper *et al.* 1991, 1993; Gibson *et al.* 1993; McNamara *et al.* 1997; Metcalfe and Buttle 2001; Laudon *et al.* 2002, 2004). While some research indicates that pre-event water is the primary source of streamflow during melt (Obradovic and Sklash 1986; Gibson *et al.* 1993; Metcalfe and Buttle 2001), other studies suggest that freshet is supplied predominantly or almost completely by meltwater inputs (Cooper *et al.* 1991, 1993; McNamara *et al.* 1997). The reason for this variability is unclear, yet issues of temporary storage (Bowling *et al.* 2003) and variable hillslope responses, particularly in areas of discontinuous permafrost (Chacho and Bredthauer 1983; Carey and Woo 2000b), may explain some uncertainty. Variability in permafrost soil properties related to ice content may also allow a broad range of responses. Furthermore the past practice of obtaining a meltwater isotopic signature from snow-cores (e.g. Cooper *et al.* 1991; Gibson *et al.* 1993) will cause a systematic error in hydrograph separations due to fractionation during melt (Taylor *et al.* 2002).

With regards to streamflow hydrochemistry, there have been few studies in a permafrost setting that investigate dissolved constituents as a method to infer runoff pathways (MacLean *et al.* 1999; Petrone *et al.* 2000; Carey and Quinton 2004). In subarctic Alaska, MacLean *et al.* (1999) and Petrone *et al.* (2000) reported that, for the summer period, permafrost-dominated catchments had higher concentrations of dissolved organic carbon (DOC), but lower concentrations and fluxes of solutes than an adjacent watershed that was nearly permafrost-free. The differences in hydrochemistry reflected different flow pathways, as the permafrost-dominated catchment restricted flow to the organic-rich active layer, enriching water in DOC. In the adjacent catchment with low permafrost distribution, water was able to infiltrate deeper into the mineral soils where DOC was adsorbed and solutes were dissolved due to greater contact time with mineral soil exchange sites. Carey (2003) reported annual DOC fluxes from the catchment utilized in this study and showed that the largest stream DOC concentration and mass flux was during freshet and derived primarily from permafrost-underlain slopes where the water table resided within the DOC-rich organic horizon. A further study by Carey and Quinton (2004) also reported how summer rainstorms lowered streamflow specific conductivity (SpC) and increased DOC; yet in combination with isotope data the dominant summer runoff pathway was identified as flow through the mineral substrate, not the surface organic layer.

Considering that most conceptual models of runoff generation in subarctic environments have been derived from hydrometric data, the objective of this study is to provide an improved and refined understanding of the sources and pathways of runoff in a discontinuous permafrost catchment during the snowmelt period utilizing hydrometric, stable isotope and hydrometric information. This study will accompany the work of Carey and Quinton (2004) who completed a complementary study for the summer rainfall-runoff period in this catchment.

Study area

Granger Basin (60°32′N, 135°18′W) is located within the Wolf Creek Research Basin, 15 km south of Whitehorse, Yukon Territory, Canada (Figure 1). The study area has a subarctic continental climate with low precipitation and a large temperature range. Mean annual January and July temperatures from the Whitehorse airport (elevation 706 m.a.s.l.) are −17.7°C and +14.1°C (1971–2000). Mean annual precipitation is 267.4 mm, of which 145 mm falls as rain, yet precipitation at the Whitehorse airport may underestimate basin precipitation by ca. 25–35% (Pomeroy and Granger 1999).

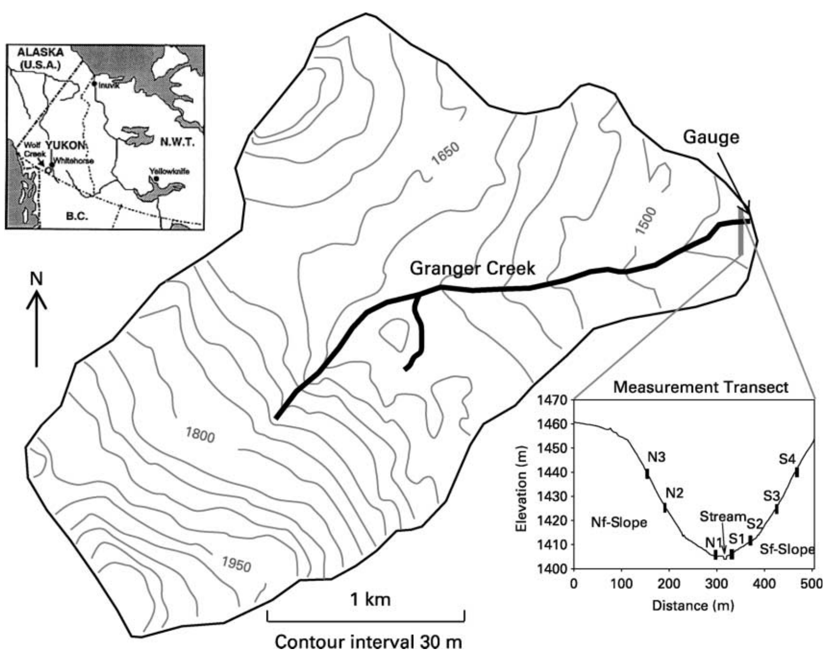

Figure 1 Study catchment (Granger Basin) within the Wolf Creek Research Basin and location of the well/tension lysimeter transects. Insets show location in Canada and position of measurement sites along transect

Granger Basin drains a ca. 6 km^2 area and ranges in elevation from 1310 m to 2250 m. The main river valley trends west to east at lower elevations, resulting in predominantly north- and south-facing slopes. The geology is primarily sedimentary comprised of limestone, sandstone, siltstone and conglomerate and is overlain by a mantle of glacial till ranging from a thin veneer to several meters in thickness. Fine textured alluvium covers most of the valley floor whereas upper elevations have shallow deposits of colluvial material with frequent bedrock outcrops present (Mougeot and Smith 1994). Permafrost is found under much of the north-facing slopes and higher elevation areas, whereas seasonal frost predominates on southerly exposures. In permafrost areas and the riparian zones, soils are capped by an organic layer up to 0.4 m thick consisting of peat, lichens, mosses, sedges and grasses. Only a few scattered white spruce (*Picea glauca*) occur within the basin, which is considered above treeline. Vegetation consists predominantly of assorted willow shrubs (*Salix*) and Labrador tea (*Ledum groenlandicum*).

Two slopes separated by the river valley were selected to compare catchment areas with and without permafrost. A permafrost-underlain north-facing (Nf-slope) slope has a gradient of ca. 0.35 and is underlain by till soils predominantly sandy in texture capped by an organic layer consisting of peat, lichens and mosses. The thickness of this organic layer varies, averaging 0.26 ± 0.10 m ($n = 30$). The active layer ranges from several decimeters to >1 m near the slope base. A south-facing slope with a ca. 0.34 gradient is underlain by seasonal frost only. Organic layer thickness is typically less (0.12 ± 0.09 m ($n = 30$)) and declines in thickness upslope from the riparian areas. Saturated hydraulic conductivity declines with depth in the organic layer and then exhibits a sharp reduction in the mineral substrate (Figure 2) (Quinton and Gray 2001). Between the slopes is a thin riparian zone that ranges from 10 m to 30 m.

Field and analytical methods

The study period is for the snowmelt period from 7 April to 6 June 2002, although intense daily measurements of all parameters did not occur until early May. Discharge was

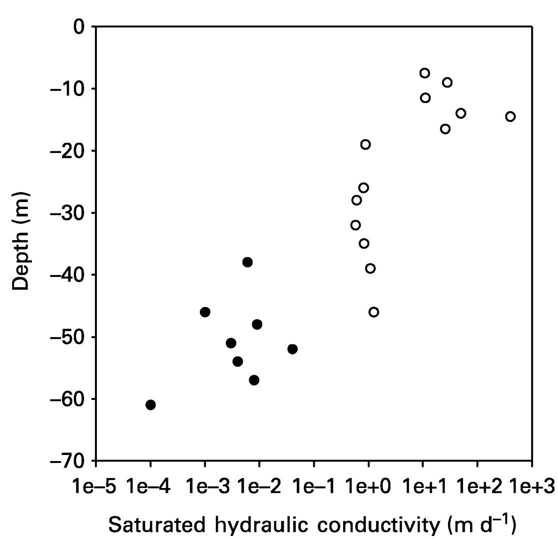

Figure 2 Saturated hydraulic conductivity of organic soils (open circles) and mineral soils (closed circles). Values for the organic soils were determined from tracer tests, whereas the mineral soil values were determined from head-recovery tests

calculated using a stage-discharge relationship for the open water period consisting of a stilling well with a float connected to an electronic logger. A stable stage-discharge relationship has existed for this catchment since 1998. During the snowmelt period, when flows are beneath ice, salt dilution was used to determine discharge (Dingman 2002). Salt dilution was continued into the open-water season whereby measurements were compared to the stage-discharge relationship and current metering. In August 2000, seven shallow groundwater wells were constructed from PVC pipe (35 mm internal diameter) with a screen section along the entire subsurface length and placed in augured holes of the same size as the pipe diameter. Three wells were placed within the permafrost-underlain Nf-slope and four within the seasonal frost S-slope. Wells were spaced along the slope length and numbered sequentially upward from the riparian zone (Figure 1). In addition, water level was monitored continuously in a well adjacent to N2 using a float attached to a 10-turn potentiometer and a datalogger.

Streamflow samples were collected several times a week until the onset of freshet. Once freshet began samples were collected at least once daily. Wells were purged and sampled a minimum of twice weekly. Snowmelt lysimeters ($0.25\,m^2$) were installed at the base of the snowpack at three locations and sampled daily at the Nf-slope and less frequently at other locations within the basin. A transect was set up across the Nf- and Sf-slopes to evaluate daily changes in snow water equivalent (SWE). Snow depth was measured at 60 points and density at 12 points on each slope using a Mount Rose snow sampler to compute mean daily slope SWE.

Isotope samples were collected in 60 ml vials sealed with airtight caps. Analyses for $\delta^{18}O$ were performed at the National Water Research Institute, Saskatoon, by CO_2 equilibrium. Accuracy of the analysis based on replicate samples was 0.05‰. Isotope compositions are expressed as δ (per mil) ratio of the sample to the Vienna standard mean ocean water standard. DOC samples were filtered through pre-combusted Whatman GF/F glass fiber filters and acidified with H_2SO_4 to 0.035 M prior to storage in sterilized 50 ml vials. Samples were kept cool until analysis, which occurred within two months of sample collection. There was no evidence of precipitate in the vials prior to analysis. Samples were run on a Technicon Autoanalyzer and DOC concentrations were determined using an automated persulfate-UV digestion with a phenolphthalein color reagent. Accuracy of DOC samples based on replicate analysis was $0.1\,mg\,C\,l^{-1}$.

A Hydrolab 4a Datasonde placed near the outlet gauge was used for continuous measurement of specific electrical conductivity (SpC), pH, stream temperature and other water quality parameters not used in this study (dissolved oxygen, oxidation–reduction potential). SpC and pH from soil waters was determined soon after collection using a portable Hydrolab Quanta sensor.

Hydrograph separation technique

Hydrograph separations are based on steady-state mass balance equations that describe the contributions of various water sources to streamflow under a simple batch mixing model with conservative tracers (Sklash and Farvolden 1979). In the case of n runoff components and $n-1$ observed tracers, $t_1, t_2, \ldots t_{n-1}$ the following n linear mixing equations can be written:

$$Q_T = Q_1 + Q_2 + \ldots + Q_n \quad (1)$$

$$c_T^{t_i} Q_T = c_1^{t_i} Q_1 + c_2^{t_i} Q_2 + \ldots + c_n^{t_i} Q_n \quad (2)$$

where Q_T is the total runoff, $Q_1, Q_2, \ldots, Q_n$ are the runoff components and $c_1^{t_i}, c_2^{t_i}, \ldots, c_n^{t_i}$ are the respective concentrations of the observed tracer t_i. The application of Equations (1) and

(2) are based on the following assumptions (Hinton *et al.* 1994; Hoeg *et al.* 2000): (1) significant differences exist between tracer concentrations of the different components, (2) tracer concentrations are constant in space and time (or any variations can be accounted for), (3) additional component contributions must be negligible, or the tracer concentrations must be similar to that of another component, (4) tracers mix conservatively, and (5) tracer concentrations of the components are not collinear.

Using Equations (1) and (2), two-component separations to evaluate the fractional contribution of a runoff component can be determined from

$$\frac{Q_1}{Q_T} = \frac{\left(c_T^{t_1} - c_1^{t_1}\right)}{\left(c_2^{t_1} - c_T^{t_1}\right)} \tag{3}$$

where the event component ($Q_1 = Q_e$) is defined as the fraction of total runoff that entered the hydrological system during the snowmelt event, whereas the pre-melt component ($Q_2 = Q_{p\text{-}m}$) is defined as the part of total runoff stored within the catchment prior to runoff. Q_2/Q_T can be solved in an analogous manner.

Concentrations of event water $\delta^{18}O$ (c_e) were calculated using the runCE method (Equation (4)) proposed by Laudon *et al.* (2002), which accounts for both the timing and the amount of water entering the soil water reservoir and the runoff of previously melted and stored water within the soil during the melt period:

$$c_e(t) = \left(\sum_{i=1}^{t} M(i)c_m(i) - \sum_{i=1}^{t} E(i)c_e(i) \Big/ \sum_{i=1}^{t} M(i) - \sum_{i=1}^{t} E(i) \right) \tag{4}$$

where $M(i)$ is the incrementally collected melwater depth, $E(i)$ is the incrementally calculated event water discharged, $c_m(i)$ is the melwater $\delta^{18}O$ obtained from lysimeters and $c_e(i)$ is the event water $\delta^{18}O$ used in the application of Equation (2). Equation (4) is then solved iteratively for each time step as $c_e(i)$ and $E(i)$ are dependent upon the calculated $c_e(t)$. Initial basin storage in the top 0.4 m (considered the hydrologically active zone) was set at 180 mm based upon fall soil-moisture profiles at three TDR sites within the basin. Change in basin storage during snowmelt was calculated by subtracting the total spring runoff from the total snowmelt and setting a constant evaporation rate of $0.17\,\mathrm{mm\,d^{-1}}$ based on previous measurements (Pomeroy *et al.* 2003). The amount of meltwater remaining in the soil after melt was determined from the difference in the amount of event water leaving the catchment. The basin-wide snowmelt rate was taken as an average between the north- and south-facing slopes. Furthermore, it was assumed that only permafrost areas (ca. 70% of the basin (Lewkowicz and Ednie 2004)) contribute to the freshet hydrograph (Carey and Woo 2001b). Further details and an application of this method can be found in Laudon *et al.* (2002, 2004).

To estimate the uncertainty associated with pre-melt ($Q_{p\text{-}m}$) and melt (Q_e) fractions of total discharge associated with systematic errors in the isotopic signatures, the Gaussian standard error method of Generaux (1998) was applied:

$$W_{fp-m} = \sqrt{\left[\frac{c_e - c_T}{\left(c_{p-m} - c_e\right)^2} W_{c_{p-m}}\right]^2 + \left[\frac{c_T - c_{p-m}}{\left(c_e - c_T\right)^2} W_{c_m}\right]^2 + \left[\frac{1}{\left(c_{p-m} - c_e\right)} W_{c_T}\right]^2} \tag{5}$$

where W is uncertainty, c is the tracer concentration, f is the mixing fraction and the subscripts $p - m$, e and T refer to the pre-melt, event and stream water components. Analytical errors were assumed equal to 0.1‰, and an error of 2‰ was used to account for spatial and temporal variability in meltwater $\delta^{18}O$ ratio and errors propagated in the calculation of c_e using the runCE method (Laudon *et al.* 2002).

Results

Snowmelt

The maximum pre-melt snow water equivalence (SWE) was measured on 19 April and yielded a SWE of 337 ± 160 mm on the Nf-slope compared with 201 ± 70 mm on the Sf-slope. Greater SWE and variability on the Nf-slope is attributed to a large drift that formed beneath the slope crest approximately 180 m upslope of the stream. Surveys of SWE showed that the Sf-slope depletes much faster than the Nf-slope, becoming snow-free by 19 May when the Nf-slope still had ca. 200 mm of SWE (Figure 3). At the end of the study period, there was only a small amount of snow remaining on the Nf-slope. Differential melt rates in alpine catchments are typical, and Pomeroy *et al.* (2003) investigated the variation in surface energetics during snowmelt for these two slopes and indicated that melt rate was controlled by both incoming energy and evolving and initial snow states.

Water samples collected from snowmelt lysimeters on the Sf-slope and Nf-slope revealed the temporal variability of $\delta^{18}O$ during the melt period (Figure 4). Meltwater $\delta^{18}O$ ratio ranged between −23.8‰ and −25.9‰ for all but the very end of the melt period when meltwater collected on the Nf-slope became significantly enriched (Figure 4). The variability in $\delta^{18}O$ may be caused by several factors, including fractionation during melt and different $\delta^{18}O$ ratios of different snow layers (Taylor *et al.* 2002). On the Nf-slope, water first appeared in the snowmelt lysimeter on 6 May and had a $\delta^{18}O$ value of −24.0‰ (Figure 4). Here, $\delta^{18}O$ in the meltwater initially became depleted as melt progressed, declining to a value of −25.8‰ on 22 May, and then gradually enriching to values between −21‰ and −22‰ at the end of the measurement period. The Nf-slope lysimeter became snow-free on 5 June, although there was still some areas of the slope that retained a snow cover. On the Sf-slope, meltwater was first collected on 24 April and had a $\delta^{18}O$ value of −23.7‰ (Figure 4) Values of meltwater $\delta^{18}O$ varied less on the Sf-slope compared with the Nf-slope, although there was a slight enrichment throughout the melt period. The last meltwater sample obtained from the Sf-slope was on 18 May.

DOC concentrations in meltwater were often below the detection limit of the analyzer. DOC in meltwater ranged from $<0.1\,\mathrm{mg\,C\,l^{-1}}$ to $1.9\,\mathrm{mg\,C\,l^{-1}}$, with a mean of

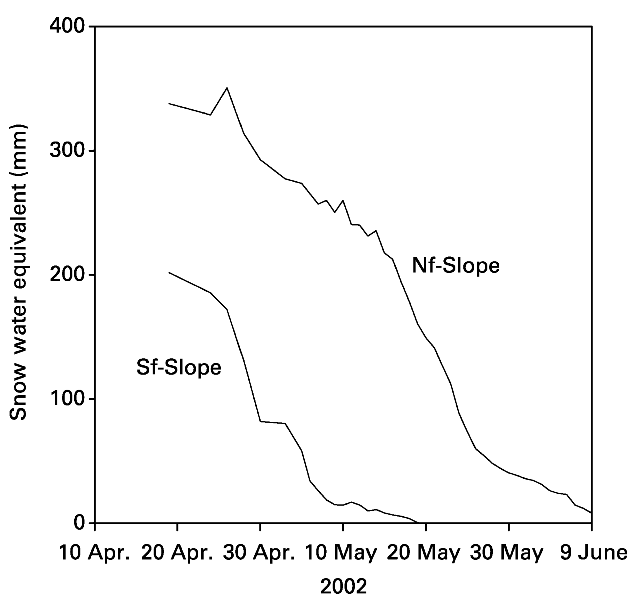

Figure 3 Snow water equivalent (mm) for the north- and south-facing study slopes during the 2002 snowmelt season

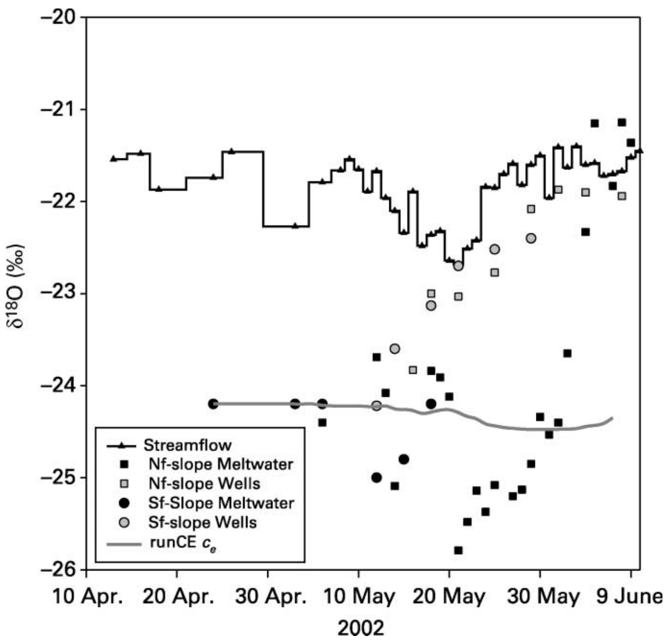

Figure 4 Time series of $\delta^{18}O$ for streamflow (triangles), $\delta^{18}O$ of event-water (c_e) calculated from Equation (4) (solid gray line), Nf-slope meltwater (black squares), water from Nf-slope well N2 (gray squares), Sf-slope meltwater (black circles) and water from Sf-slope well S2 (gray circles)

$0.24\,mg\,C\,l^{-1}$. The variability in snowmelt DOC may in part be due to the density of vegetative cover over the lysimeter, as highest values of DOC concentrations were recorded at a melt lysimeter at the base of the Nf-facing slope where considerable shrub vegetation existed. There was no temporal trend in DOC concentration in snowmelt waters. Snowmelt specific conductivity was low, averaging $4.4 \pm 6\,\mu s\,cm^{-1} (n = 46)$ for the study period. While there were three samples that had SpC values $> 10\,\mu s\,cm^{-1}$, most meltwater samples were dilute with respect to dissolved ions. There were no apparent temporal or spatial trends in meltwater SpC.

Subsurface waters

Water table measurements began when wells became snow-free and did not have ice obstructions which could interfere with sampling (Figure 5a, b). Within the Nf-slope, water tables were at a similar near-surface position regardless of slope position (Figure 5a). There was a gradual lowering of water level throughout the melt period as the water table declined atop the descending frost table. Within the Sf-slope, sites S1, S2 and S4 had consistent yet consecutively lower water table levels away from the stream (Figure 5b). Site S3 had near-surface water tables for the majority of the melt period, yet levels declined dramatically after 28 May, which may have been the result of seasonal ground thaw at depth.

Values of $\delta^{18}O$ sampled at N2 and S2 became gradually enriched throughout melt (Figure 4). Initial meltwater samples from wells were depleted in $\delta^{18}O$ compared with streamflow, suggesting that most of the liquid water within the wells at the beginning of sampling was derived from snowmelt infiltration. As the melt period progressed and soil thawing increased, $\delta^{18}O$ enriched steadily from values $> -24‰$ to values approaching streamflow near $-22‰$. There was no significant difference (Pearson product moment correlation at the 95% confidence level) between the Nf-slope and Sf-slope, suggesting that

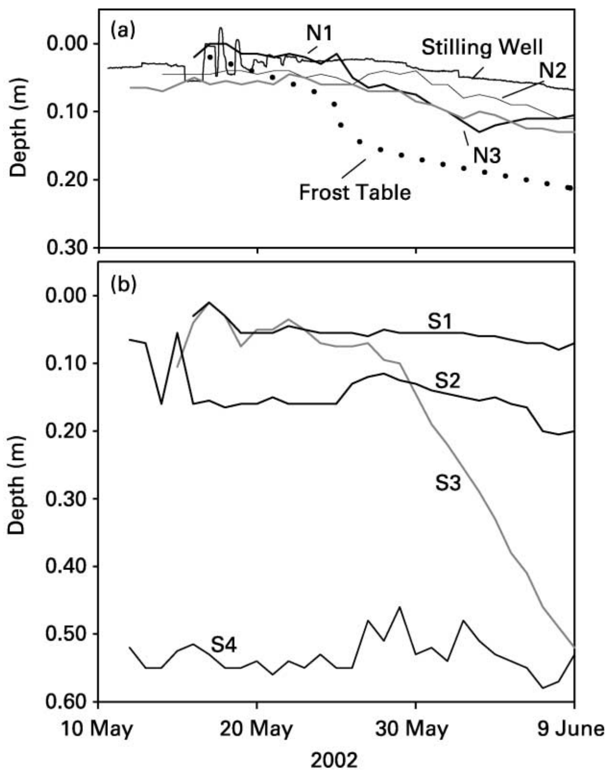

Figure 5 Water table position (m) during melt for (a) north-facing slope wells underlain with permafrost and (b) south-facing slope wells with seasonal frost only. Manual measurements were taken between 13:00 and 14:00

the catchment is well mixed with regards to the melt isotopic signature. DOC concentrations declined in groundwater sampled at all wells during melt (Table 1). Within the Sf-slope, concentrations fell from $25.8 \pm 9.7\,\mathrm{mg\,C\,l^{-1}}$ between 10 and 15 May to $7.8 \pm 4.4\,\mathrm{mg\,C\,l^{-1}}$ between 1 and 5 June. For the Nf-slope, DOC concentrations were higher throughout the

Table 1 Dissolved organic carbon concentration from groundwater wells obtained during the snowmelt period, 2002. N are north-facing wells underlain with permafrost and S are south-facing wells with seasonal frost only

	DOC concentration (mg C L^{-1})						
Date	**N1**	**N2**	**N3**	**S1**	**S2**	**S3**	**S4**
9-May-02				33.1			
11-May-02					35.5		
12-May-02							
14-May-02							
15-May-02					21.4	15.5	
16-May-02	78.9			27.2			
18-May-02	50.1	37.3	42.7	18.9		20.1	7.3
21-May-02	30.9	22.0	36.4	13.4		15.8	7.7
25-May-02	20.9	16.6	22.6	10.9		17.5	5.6
29-May-02	13.9	16.5	20.2	5.3	23.8	13.1	3.1
1-June-02	17.7	22.7	16.2	5.3		10.1	5.1
4-June-02	13.2	22.7	19.5	4.7	15.6	9.1	4.5

duration of melt, averaging $55.4 \pm 21.3\,\mathrm{mg\,C\,l^{-1}}$ between 15 and 20 May and declining to $19.1 \pm 4.5\,\mathrm{mg\,C\,l^{-1}}$ between 1 and 5 June despite water tables remaining near the surface. There was some variability with topographic position on the Sf-slope, with the riparian zone and upslope locations having lower DOC levels, yet this pattern was not apparent within the permafrost-underlain Nf-slope.

Streamflow

The 2002 snowmelt freshet period lasted from 13 May to 1 June (Figure 6a). Prior to the onset of freshet, flows were at their lowest annual level at ca. $10\,\mathrm{l\,s^{-1}}$. On 13 May, discharge slowly began to rise until midday on 16 May when a sudden increase occurred. Discharge again rose abruptly to ca. $400\,\mathrm{l\,s^{-1}}$ on 17 May and stayed at that approximate level until 24 May when a rapid increase to a peak freshet value of $1480\,\mathrm{l\,s^{-1}}$ occurred over 8 hours. Following this peak, daily minimum flows declined gradually until 1 June. Superimposed atop the seasonal melt pattern are large daily cycles caused by the diurnal variability in snowmelt. Following 1 June, daily melt cycles were small and flows returned to post-freshet summer levels.

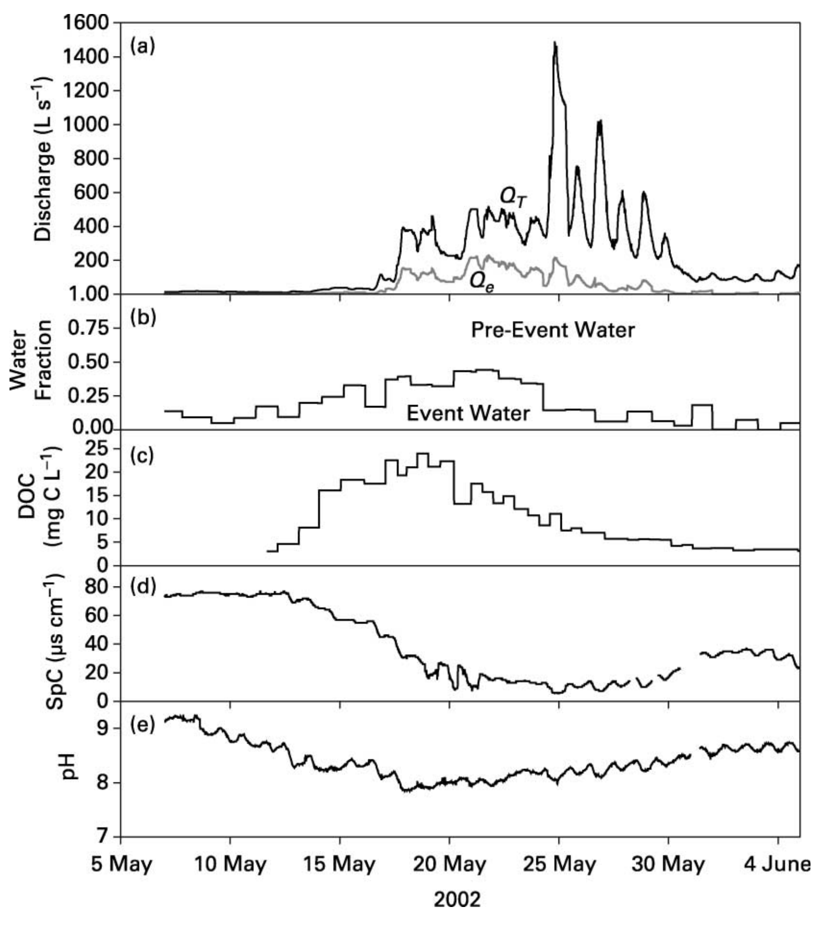

Figure 6 Time series of (a) total streamflow discharge (solid line) and event water contribution to streamflow discharge based on 2-component hydrograph separation (gray line), (b) event-water and pre-event-water fraction based on two-component hydrograph separation, (c) streamflow dissolved organic carbon, (d) streamflow specific conductivity and (e) streamflow pH

Prior to the increases in flow, streamflow $\delta^{18}O$ was between $-22‰$ and $-21‰$ (Figure 4), which were similar to baseflow levels in the summer of 2001 (Carey and Quinton 2004). Beginning on 13 May, $\delta^{18}O$ declined slowly over 9 days, reaching a seasonal minimum of $-22.7‰$ on 22 May. Following this, streamflow became slowly enriched and then, on 24 May, stream $\delta^{18}O$ rose abruptly to $-21.8‰$ coinciding with the rapid rise to freshet. For the remainder of the melt period, streamflow $\delta^{18}O$ was $> -22‰$. Stream water quality parameters showed strong dilution/acidification trends during freshet (Figure 6c–e). Prior to the onset of freshet on 13 May, DOC concentrations remained stable between 2 and $3\,mg\,C\,l^{1}$ (Figure 6c) After streamflow began to slowly increase, DOC concentrations rose rapidly to a maximum of $24\,mg\,C\,l^{-1}$ on 18 May, six days before peak discharge. Following this DOC peak, concentrations declined exponentially, reaching pre-freshet levels in early June after which DOC responded much less to changes in discharge. The relation between DOC and discharge exhibited large hysteresis during the melt period as values on the recession limb were significantly greater than those on the receding limb at equivalent discharges. Specific conductance exhibited a strong dilution trend, with pre-event values at ca. $80\,\mu s\,cm^{-1}$ declining gradually throughout freshet to values below $10\,\mu s\,cm^{-1}$ (Figure 6d). The lowest SpC value observed was $6\,\mu s\,cm^{-1}$ and was associated with the peak freshet event on 24 June. Following the snowmelt period SpC gradually increased towards summer values, which fluctuated around $40\,\mu s\,cm^{-1}$ for 2002. Streamflow pH declined from values >9 to <8 over the course of the melt period (Figure 6e). A strong daily cycle was evident in the pH signal corresponding to lower pH during times of maximum runoff from hillslopes as biological activity was negligible within the stream at this time. Following melt, pH slowly increased to values ca. 8 for the remainder of the year.

Hydrograph separations

For hydrograph separation, concentrations of melt-water $\delta^{18}O$ (c_e) were determined from Equation (4) using: (i) daily average $\delta^{18}O$ meltwater inputs (c_m) from both slopes, (ii) a basin-wide meltrate taken as the average from both slopes, and (iii) only the permafrost-underlain areas of the catchment (ca. 70%) contribute to streamflow. Hydrometrically, during the 7 May–5 June freshet period, 258 mm of snowmelt water entered the catchment and 110 mm of water exited, of which 105 mm was streamflow. For the study period, c_e showed little variability compared with c_m measured in lysimeters, decreasing from $-24.2‰$ at the onset of melt to a depleted value of $-24.48‰$ on 27 May. Values of c_e then began to increase as meltwater became enriched and streamflow began draining the significant storage within the hillslope reservoir. Concentrations of pre-melt water stored within the catchment ($c_{p\text{-}m}$) was taken as $-21.6‰$, which represents the baseflow value taken prior to any melt, and is also similar to baseflow and groundwater values at the end of 2001 (Carey and Quinton 2004).

Results of the two-component hydrograph separation using $\delta^{18}O$ provided a total event water contribution of ca. 21% for the 7 May–5 June freshet period. The relative fraction of event water was $<20\%$ at the onset of melt and rose steadily as flows increased (Figure 6b). The event-water contribution rose to a peak value of 44% on 21 May and remained $>30\%$ until the onset of freshet on 24 May when the event-water fraction declined dramatically to ca. 15%. Following this decline, the event-water fraction exhibited some variability and declined slowly until the end of the snowmelt period. Standard error was calculated using Equation (5) based upon analytical uncertainties and ranges between 7–12% for all periods of the separation.

Discussion

Snowmelt is the most important hydrological and hydrochemical event in subarctic and arctic catchments, transferring approximately half of the annual precipitation to the stream over a period of several weeks (Woo 1986). As observed in Granger basin, this rapid delivery of meltwater imparts a distinct hydrochemical signature to the streamflow. In turn, this hydrochemical signal can be used to infer hydrological pathways and assess the role of seasonal frost and permafrost in controlling catchment dynamics.

Approximately 90% of SWE on the Sf-slope and 30% of SWE on the Nf-slope melted prior to the onset of freshet on 13 May (Figure 6a). This large delivery of meltwater into Sf-slope soils without a corresponding rise in streamflow suggests that this permafrost-free slope does not have a significant hydrological connection to the stream during this time. Conversely, on the Nf-slope, the timing of the main melt period (Figure 3) corresponds closely with the timing of streamflow freshet. Within the Sf-slope, the riparian zone saturated quickly, while upslope areas had increased infiltration capacities and deeper water tables as unsaturated seasonally frozen soils allow vertical infiltration. Conversely, within the Nf-slope, this water table was perched above the mineral substrate within the organic layer at all sites.

The decline in SpC on 13 May coincides with the initial rise in streamflow. As SpC of meltwater changes with soil contact time, often dramatically during the initial contact period with mineral soils (Pilgrim *et al.* 1979), the decline in streamflow SpC indicates that meltwater is initially reaching the stream through near-surface pathways with little interaction with the mineral substrate. After 13 May, streamwater becomes increasingly diluted with regards to dissolved ions, falling to values nearer to meltwater than mineral layer water (SpC ca. 80 $\mu s\,cm^{-1}$, Carey and Quinton (2004)). The concomitant rise in DOC indicates that water is flowing through porous near-surface organic layers that are highly enriched in DOC (Table 1). The decline in DOC within hillslope wells and streamflow, despite the continued rise in the freshet hydrograph, occurs due to flushing of this limited source of labile carbon from the organic layer (Carey 2003). This combined evidence of declining SpC, increasing DOC and slope melt patterns suggests that the vast majority of freshet water originates from permafrost-underlain areas, an observation that has been previously reported in other subarctic settings (Chacho and Bredthauer 1983; Carey and Woo 1998 As meltwater on the Sf-slope has infiltrated the seasonally frozen mineral soils prior to freshet, coming into contact with mineral exchange sites and increasing SpC, it is reasonable to assume that this slope contributes only a small portion of water to streamflow as there was no SpC change even when streamflow decreased despite most of the snowpack melting.

Hydrograph separations

Due to the typical depletion in meltwater $\delta^{18}O$, two-component hydrograph separation is a useful tool to evaluate the relative contributions of meltwater and water that previously existed within the catchment (Laudon *et al.* 2002; Taylor *et al.* 2002). Despite this, there is considerable variability in the timing and magnitude of meltwater delivery across Granger basin, and results must be viewed with some caution considering this variability and the simplifying assumptions used in this study. At the onset of freshet, event-water contributions (Q_e) slowly began to rise from $<20\%$ to ca. 25% between 16–23 May. This rise coincided with a warm rapid melt period on the Nf-slope (Figure 3). When viewed with respect to the $\delta^{18}O$ of subsurface water (Figure 4), the initial depletion in soil $\delta^{18}O$ is caused by meltwater that has infiltrated prior to 13 May and has been residing in temporary storage within the organic layer (becoming enriched in DOC) prior to drainage pathways becoming unfrozen, allowing rapid drainage to the stream. Prior to 13 May, approximately 137 mm of snow had melted while only 5 mm had run off and evaporated. The subsequent depletion of the streamwater $\delta^{18}O$ and enrichment of soil $\delta^{18}O$ (Figure 4) suggests that this is the initial breakthrough of stored meltwater which sustains high event-water fractions for approximately 7 days. During this period when the stored meltwater drains, hillslope $\delta^{18}O$

becomes enriched to values that reflect a more mixed signal of melt and pre-melt pore waters. Following this, Q_e declines despite continued increases in the stream hydrograph, falling to levels below ca. 10% near the end of the melt period. It is hypothesized that, at this time, much of the basin is snow-free and soil thawing is rapid, allowing additional inputs of meltwater to mix with thawing soil water and displacing pre-melt soil water, increasing the value of $Q_{p\text{-}m}$. By the end of the melt period in June, Q_e has declined to only a small fraction of total flow Q_T.

Uncertainties in melt and pre-melt fractions of total streamflow are associated with systematic errors in the isotope signatures, and the estimated accuracies for separation components (7–12%) are similar to values reported by other workers during snowmelt (Metcalfe and Buttle 2001; Laudon *et al.* 2004). The random errors associated with spatial variability are difficulty to quantify and the assumption that only two reservoirs exist within a catchment of this scale may be open to question. There have been relatively few and conflicting reports of hydrograph separation in a permafrost setting during melt (Obradovic and Sklash 1986; Cooper *et al.* 1991, 1993; Gibson *et al.* 1993; McNamara *et al.* 1997; Metcalfe and Buttle 2001) and results from this study indicate that the bulk of meltwater delivered to the stream over the course of freshet is pre-melt water that existed within the catchment prior to freshet. Despite the anecdotal observations that large pre-melt contributions are unlikely considering the frozen ground, in the subarctic, meltwater is able to infiltrate, mix and displace pre-event waters, particularly in the organic layer of permafrost-underlain slopes. To the authors' knowledge however, the pattern of greater event-water contribution during the early stages and pre-melt in the later stages has not previously been reported.

Snowmelt runoff mechanisms

Previous conceptual models of snowmelt runoff generation in the subarctic have been primarily based on hydrometric data (Carey and Woo 2001a; Dingman 1971; Santeford 1979; Slaughter and Kane 1979). A common theme observed across subarctic North America and one that is confirmed in this study is that there are two major controls on snowmelt runoff: (1) permafrost distribution defines basin contributing areas as lateral flow is confined to permafrost-underlain slopes due to their ability to restrict deep percolation and (2) surface organic soils play a key role in rapidly conveying water to the stream due to their unique hydraulic properties (Quinton *et al.* 2000), controlling the timing and magnitude, and characteristics of the freshet hydrograph. Surface runoff is rarely observed due to the high infiltration and storage capacity of soils, yet preferential pathways such as inter-hummock (Quinton and Marsh 1998) and pipeflow (Gibson *et al.* 1993; Carey and Woo 2000) may be important. Although no preferential flow was observed, results from this study using stable isotope and simple hydrochemical indicators support these observations and highlight the increased importance of temporary storage at the onset of melt.

At the beginning of the melt period, meltwater percolating from the snowpack infiltrates the frozen ground, which is typically unsaturated. On slopes with permafrost soils and an organic cover, infiltration is restricted to the porous organic soils whereas, on slopes with seasonal frost, percolation is uninhibited unless there are ice-rich layers at depth (Kane and Stein 1983). The isotopic signature of liquid soil water at this time becomes strongly depleted with respect to $\delta^{18}O$ suggesting large meltwater recharge (Figure 4). Additionally, soil temperatures are ca. 0°C, as meltwater infiltrating the soil provides little sensible heat to melt pore water, thus reducing the mixing of melt and pre-melt pore water until soil thaw begins. During these early melt stages, there is little lateral transfer of water to the stream until the unsaturated storage capacity of the soils declines and soil thaw allows runoff pathways to become exposed. Once snow-free areas increase and a perched water table is formed, runoff is rapidly conveyed to the stream from permafrost-underlain slopes with a strong meltwater isotope signature (Figure 4). Hydrochemically, DOC that is removed from organic soils in the meltwater solution is flushed during this initial runoff pulse (Carey 2003). The decline in streamflow SpC occurs following the initial DOC and $\delta^{18}O$ depletion signal from the initial storage release, as meltwater is then allowed to pass rapidly from the melting snowpack to the stream via the organic layer, bypassing the significant interaction with underlying mineral soils. The fraction of meltwater contributing to streamflow declines, however, as pore water within the organic layer melts, mixing with and being displaced by meltwater. During peak freshet, the continued dominance of flow through the organic layer is supported by the decline in stream Spc and pH, the latter occuring due to acidification from organic-layer water. Near the end of the freshet period, meltwater infiltration continues, yet mixing with subsurface soils, including those from the mineral layer, dilutes the melt $\delta^{18}O$ signature and displaces pre-melt stored water. As the freshet period ends, thaw allows interaction with the mineral soils as evidenced by the increase in stream SpC and pH whereas DOC has declined to near pre-freshet levels.

Summary

Stable isotope ($\delta^{18}O$) and hydrochemical (DOC, SpC, pH) data were used in combination with hydrometric data to evaluate the sources and pathways of snowmelt runoff generation in a mountainous subarctic catchment with discontinuous permafrost. Results from hydrograph separations indicate that meltwater accounted for ca. 20% of freshet water, with the vast majority of streamflow being derived from soil water displaced through the organic layer to the stream during melt. Areas with southern exposures that were permafrost-free had only a limited contribution to freshet. On permafrost slopes, meltwater initially infiltrated porous organic soils, forming a saturated layer perched atop the mineral–organic layer interface. As melt progressed and bare ground became exposed, sufficient thaw occurred to allow drainage from the organic horizon and water was rapidly conveyed to the stream with an initial isotopic and hydrochemical signature of meltwater passing through the organic layer. Near the end of the melt, meltwater contributions to streamflow declined despite increases in flow as the water table on the Nf-slope declined to deeper depths, allowing increased mixing between melt and pre-melt hillslope waters.

Despite the apparent conclusion from a variety of data sources that permafrost areas contribute to the melt hydrograph whereas areas of seasonal frost do not, there is a significant degree of spatial variability in a catchment of this scale and it is uncertain whether all runoff pathways and sources of water have been accurately identified. Further integrated studies are required to establish a better understanding of hillslope processes in this environment and how they are applicable to other areas of the subarctic.

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