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Hydrologic functions of wetlands in a discontinuous permafrost basin indicated by isotopic and chemical signatures

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Abstract

The hydrology of subarctic, discontinuous permafrost regions is sensitive to the effects of climatic warming, because pronounced changes in water storage and runoff pathways could occur with small additional ground heating. The objective of this study is to understand the hydrologic functions of unique land-cover types in this region (channel fens, flat bogs, and peat plateaus) using isotopic and chemical signatures of surface and subsurface water, as well as hydrometric measurements. The study was conducted in a 152-km^2 basin of Scotty Creek, located in the central part of the Mckenzie River basin in northern Canada. The headwater of Scotty Creek, Goose Lake had a strongly enriched isotopic composition due to evaporation. The stream water composition changed downstream, as the lateral drainage from the active layer of peat plateaus contributed isotopically light and chemically dilute water to channel fens that are part of the drainage network. Flat bogs received drainage from peat plateau in addition to direct precipitation, and were internally drained or drained water to adjacent channel fens. Average evapotranspiration estimated from the chloride-balance method was 280–300 mm/yr, which was consistent with the hydrometric estimate (precipitation minus runoff) of 275 mm/yr indicating a potential applicability of this method to ungauged basins. Tracer-based hydrograph separation showed that the direct snowmelt contribution to spring runoff was less than half of total discharge, suggesting an importance of the water stored over winter in lakes and wetlands. The total amount of water stored over winter in the basin was estimated to be 140–240 mm, which was comparable to the average annual basin discharge (149 mm).

Keywords: Chloride; Oxygen-18; Deuterium; Snowmelt; Wetland; Permafrost; Active layer

1. Introduction

The Mackenzie River basin in northern Canada is experiencing some of the greatest warming in the world over the last few decades, and the effect of this warming on the hydrological regime of the region is of major concern (Stewart et al., 1998). The Mackenzie River, with a drainage basin of 1.8×10^6 km^2, provides an important water input to the Arctic Ocean, thereby contributing to the thermohaline circulation of the world's oceans, which regulates the global climate (Carmack, 2000).

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Therefore, there is a critical need to improve the understanding of hydrological processes throughout the north, including its many biophysical facets from the northern reaches of the boreal forest to the open tundra at higher latitudes (Rouse, 2000).

The central part of the Mackenzie River basin, drained by the lower Liard River, is located in the Continental High Boreal wetland region of Canada (NWWG, 1988) and in the zone of discontinuous permafrost (Hegginbottom and Radburn, 1992). River basins in the region are characterized by extensive, flat headwater zones with a high density of open water and wetlands, with marginally steeper regions in the lower drainage network where confined channels are incised into mineral soils (Craig, 1991; Gibson et al., 1993b). Headwater regions are dominated by a mosaic of peat plateaus underlain by permafrost, and fens and bogs that are only seasonally frozen (Robinson and Moore, 2000). The lateral exchange of water and energy between peat plateaus and wetlands, and the storage and transmission of water through the network of wetlands are unique hydrological aspects of this region. The hydrology of the discontinuous permafrost terrain is believed to be particularly sensitive to the effects of climatic warming, because pronounced changes in water storage and runoff pathways could occur with small additional ground heating (Rouse, 2000). This has an important consequence in relation to the global carbon cycle, because the rates of accumulation and decomposition of organic carbon in northern peat lands, a major fraction of the world's carbon pool, are sensitive to the hydrologic regime (Robinson and Moore, 1999).

Evaluating the feedback between climate and hydrology of northern basins will require a coupled atmosphere-hydrological-land surface models that incorporate key processes affecting the water and energy cycles involving permafrost (Stewart et al., 1998). The first attempt to simulate snowmelt runoff in the lower Liard region was made by Pietroniro et al. (1996), who found that the model parameters based on studies in temperate regions did not adequately transfer to this region. They emphasized the need to understand and properly represent the unique hydrological processes in this discontinuous permafrost region, particularly the hydrologic functions of wetlands. Wetlands store runoff water and attenuate peak flow if they have sufficient storage capacity. However, when the storage capacity is exceeded, individual wetlands become connected to form a flow system (Price and Maloney, 1994; Quinton and Roulet, 1998). Storage capacity is generally smallest in spring when the snowmelt water supply exceeds the storage capacity of the frozen, saturated wetland soils, but increases with time as the soil thaws and water is lost through seepage and evaporation (Roulet and Woo, 1986). In low-gradient arctic and subarctic regions, the connectivity of wetlands over entire drainage basins ($10^2 - 10^3$ km^2) is relatively high during spring, but decrease in summer as demonstrated by Bowling et al. (2003) using remote sensing techniques. Pietroniro et al. (1996) and Quinton et al. (2003) also demonstrated the utility of remote sensing techniques in the analysis of drainage pathways and response of northern basins, where accessibility is severely limited.

Environmental tracer methods also offer an alternative approach that can complement labour-intensive hydrological process studies (Gibson, 2001). The time-series of tracer concentrations at a basin outlet, combined with other hydrometric information, can be used to quantify the integrated effects of hydrologic processes in the entire basin. For example, mass balance of conservative tracers, such as chloride, has been used to estimate basin-scale evapotranspiration averaged over a long time period (Claassen and Halm, 1996). Isotope tracers are commonly used to evaluate the relative contribution of difference source waters during storm runoff (Sklash and Farvolden, 1979). Previous field studies in the Manners Creek basin (300 km^2) used oxygen-18 and deuterium tracers to examine runoff generation mechanisms (Gibson et al., 1993b) and quantify evaporation (Gibson et al., 1993a). The successful application of these large-scale tracer methods, however, requires an understanding of smaller-scale processes. For example, it is commonly assumed that in-channel storage of water has negligible effects in the tracer-based hydrograph separation. This assumption will likely be invalid in wetland-dominated regions, since compared with confined-channel systems, the amount of water draining through wetlands is small relative to the amount of water in storage, and the residence time of this drainage water is relatively large.

The present study examines the chemical and isotopic compositions of surface and subsurface waters in a 152-km^2 drainage basin in the lower Liard River valley. The objectives are to (1) identify the chemical and isotopic characteristics of water in fen, flat bog, and peat plateau and examine the hydrological connection among these land-cover types, (2) assess the applicability of the chloride mass balance method to estimate evapotranspiration in northern basins, (3) evaluate the relative contribution of snowmelt water during spring runoff, and (4) examine the snowmelt storage capacity of wetlands. Four different tracers (chloride, electrical conductivity (EC), oxygen-18, and deuterium) are used in this study, combined with standard hydrometric measurements. Using multiple tracers provides complementary information that can be used to check the consistency of results and increase the confidence in data interpretation (Mazor, 1976).

2. Study site

The study was conducted at Scotty Creek (61°18′N, 121°18′W) located in the lower Liard River valley, 50-km south of Fort Simpson, Northwest Territories, Canada (Fig. 1a). The basin is typical of the numerous wetland-dominated, discontinuous permafrost basins in this northern region, where interconnected wetlands store and transmit runoff water received from peat-covered uplands. The stratigraphy in this region includes an organic layer of varying thickness (less than 0.5–8 m) overlying a silt–sand layer, below which lies a thick clay to silt–clay deposit of low permeability (Aylesworth and Kettles, 2000). The 1971–2000 mean annual air temperature in Fort Simpson was −3.2 °C, and the mean annual precipitation was 369 mm, of which 170 mm was snow (MSC, 2002a). The average temperature and precipitation of the last five years (1997–2001) was −1.5 °C and 432 mm, respectively (MSC, 2002a). The difference between the 1971–2000 and 1997–2001 periods was mostly due to warm winters and wet summers. Snowmelt usually commences in late March and the stream discharge starts to increase in the mid April. Normally, only small amounts of snow remain on the ground by May (Hamlin et al., 1998).

Scotty Creek has a drainage basin of 152 km^2, and contains the major ground cover types within the region, including channel fens, flat bogs, peat plateaus, and other wooded uplands. Elevation ranges between 240 and 290 m. Quinton et al. (2003) estimated that approximately 20% of the Scotty Basin is covered by channel fens and 10% is covered by flat bogs, which is similar to other drainage basins in this region. Scotty Creek has a relatively low drainage density (0.016 km km^{-2}) and basin slope (0.0032) compared to other basins examined by

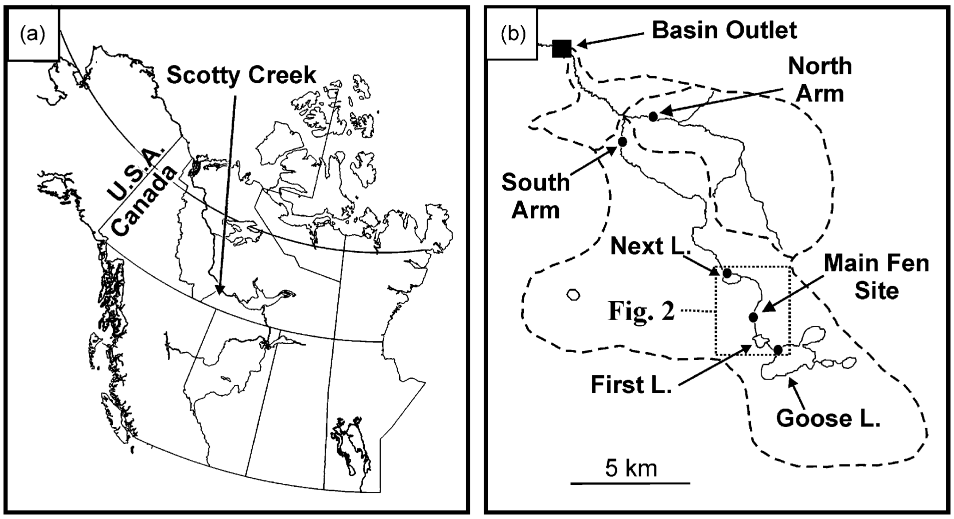

Fig. 1. Site map showing Scotty Creek and study nodes (after Quinton et al., 2003).

Quinton et al. (2003). The major drainage system of Scotty Creek is composed of channel fens, open stream channels, and intervening lakes (Fig. 1b). Open stream flow predominates over the lower one-third of both the North and South Arms, as well between the confluence and the basin outlet. The remaining, upstream portions of both arms are composed of channel fens, with intervening lakes. Fig. 2 shows a classified satellite image (Quinton et al., 2003) of the area between Goose Lake and Next Lake, where interconnected channel fens provide drainage pathways, a major hydrological characteristic of this region.

The channel fens are located along the drainage network of basins, often taking the form of broad, 50– > 100 m wide channels (Fig. 2). The arrangement of the channel fens on the landscape, and observations of flow over their surfaces suggests that their hydrological function is primarily one of lateral flow conveyance. The surface of these channel fens is composed of a buoyant mat of *Sphagnum riparium*-dominated peat, approximately 0.5–1.0 m in thickness. Field measurements at Main Fen (Fig. 2) indicated that below the buoyant mat was a denser organic layer, and that mineral soil occurred at a depth of 3 m below the water surface. The surface of the buoyant mat supports sedges and other emergent, aquatic vegetation including various herbs and shrubs. Although this surface rises and subsides with the water surface, in many places it remains approximately 5–20 cm below the water surface. Therefore, surface flow, although often diffuse, is a major runoff mechanism in channel fens. The surfaces of flat bogs are relatively fixed, and are covered with *Sphagnum* species, overlying yellowish peat with well-defined *Sphagnum* remains (Zoltai and Vitt, 1995). In the flat bogs and peat plateaus, where the water table is usually below the ground surface, subsurface flow is the predominant lateral flow path (Quinton and Marsh, 1999).

Fig. 3 shows the working conceptual model of hydrological processes in this region. Peat plateaus are underlain by permafrost, and have an active layer thickness of between 0.8 and 1 m. Fig. 3 depicts the condition in late summer with the fully thawed active layer. The frost table rises to the surface in winter and the soil becomes completely frozen. The permafrost is absent under channel fens and flat bogs. The mineral soil occurs at depths of several meters. The permafrost thickness has not been measured in the study site, but Burgess and Smith (2000) reported 5–10 m in the boreholes under peat-covered sites located near Fort Simpson. Therefore, the permafrost is expected to penetrate into the mineral soil. The ground surfaces of

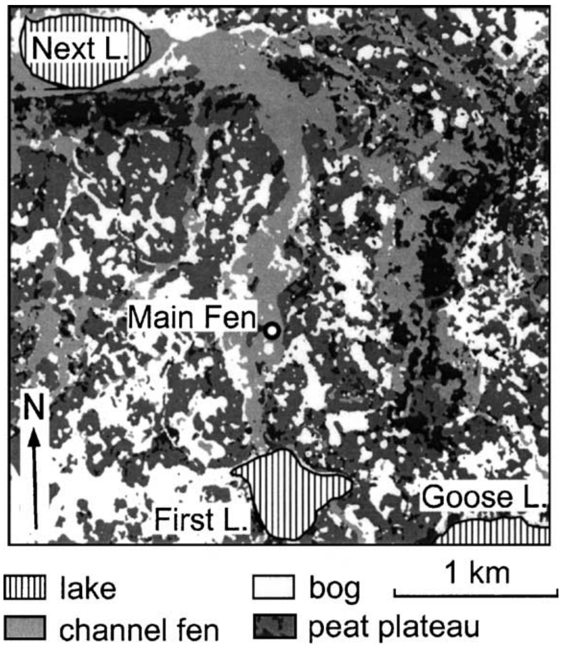

Fig. 2. Distribution of lakes and channel fens forming the integrated drainage system of Scotty Creek based on a classified image of IKONOS satellite (after Quinton et al., 2003).

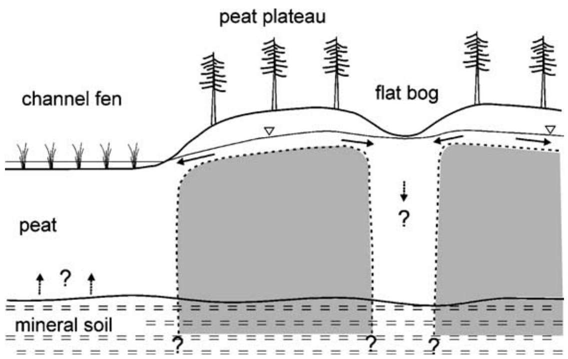

Fig. 3. A schematic cross-section of a peat plateau, and adjacent channel fen and flat bog areas with arrows indicating subsurface flow. Shaded areas indicate the frozen soil. Note that the vertical dimension is greatly exaggerated and plants are not drawn to scale. The depth to mineral soil is 3–4 m at the Main Fen site.

peat plateaus rise 1–2 m above the surrounding bogs and fens. Mature plateaus support shrubs and trees (*Picea mariana*), with the ground cover composed of lichens and mosses overlying sylvic peat containing dark, woody material, and the remains of lichen, rootlets and needles. The surfaces of peat plateaus are relatively dry, and the flow of subsurface water is normally restricted to the saturated zone between the water table and the frost table. The vertical position of the saturated zone strongly influences the rate of lateral subsurface drainage owing to the substantial decrease in permeability with depth (Quinton et al., 2000; Hoag and Price, 1997).

The literature from warmer regions suggests that fens receive groundwater discharge and bogs recharge groundwater (Siegel and Glaser, 1987; McNamara et al., 1992), but the hydrogeological functions of channels fens and flat bogs in the discontinuous permafrost region have not been well documented. The frozen soils have low permeability (Woo and Winter, 1993), which restricts the subsurface flow. As a result a large amount of snowmelt runoff is generated on peat plateaus in early spring and collects in flat bogs and channel fens. Flat bogs have been observed to drain into channel fens during snowmelt, and thereby contribute to the basin runoff. However, the spatial and temporal variation of such a hydrological connection between the bogs and channel fens, and their influence on basin discharge remain poorly understood.

3. Methods

3.1. Hydrometric measurements

Stream discharge was monitored continuously by the Water Survey of Canada (WSC) at the basin outlet (Fig. 1b). During the summer in 1999 and 2000, water level was monitored at nodes along the main drainage system at Goose Lake, Main Fen site, Next Lake, and near the confluence of the South and North Arms of Scotty Creek (Fig. 1b) with pressure transducers (Druck PDCR 950) installed in a slotted 5-cm diameter PVC stilling well and connected to data loggers (Campbell Scientific 21X). On a peat plateau adjacent to Main Fen, water-table wells made of slotted 5-cm PVC pipes were installed at two locations, 2.7- and 14.2-m away from the edge of the channel fen. The bottom of these wells was within the active layer, 50–60 cm below the surface. A 5-cm inner diameter PVC piezometer was driven into the mineral soil under Main Fen. The piezometer had a 20-cm long slotted screen centred at a depth of 3.0 m and a cone-shaped drive point at the bottom. Undisturbed soil samples were collected from various depths in soil pits to determine volumetric water content and porosity. Soil temperature was measured by a string of thermistors connected to data loggers on the peat plateau adjacent to Main Fen. At the same location snow depth was measured by a sonic ranger (Campbell Scientific SR50), and liquid water content of the soil was measured by a water content reflectometer (Campbell Scientific CS615) calibrated to the local soil (Quinton et al., in review).

Rainfall was recorded at the Jean-Marie gauging station, located 12-km east of the Scotty outlet by the WSC in 1999 and 2001. This station did not record the data in 2000 due to equipment failure. Cumulative rainfall over a period of May 1–September 30, 1999 was 211 mm at Jean-Marie and 238 mm at Fort Simpson. The May–September rainfall in 2001 was 346 mm at Jean-Marie and 316 mm at Fort Simpson (MSC, 2002a). The cumulative rainfall was reasonably similar between the two stations, although the rainfall depth of individual storms often varied substantially. Therefore, annual rainfall data at Fort Simpson was used in the water balance analysis. The Fort Simpson meteorological station uses a Nipher snow gauge for winter precipitation, which typically has wind-undercatch errors of 0–25% depending on wind speed (Goodison et al., 1998). Uncertainty in cumulative precipitation due to instrumental error is expected to be 10% in winter and less than 5% in summer. Daily mean air temperature at Fort Simpson was estimated from daily maximum and minimum temperatures (MSC, 2002a). Annual snow surveys were conducted on March 23–30, 1999 (Carter and Onclin, 1999), March 30–April 1, 2000 (Onclin et al., 2000), and March 23–25, 2002 (Onclin and Best, 2002) at over 50 locations in the Fort Simpson area (Hamlin et al., 1998). Snow surveys were not conducted in 2001. The data from 25 locations within 40-km of Scotty Creek were used in this study to estimate the average snow water equivalent in the area.

3.2. Water and snow sampling and analysis

Surface water samples were collected at the basin outlet and the nodes (Fig. 1b). Sampling interval varied from a week to a few months during the study period. Subsurface water in the active layer was collected from soil pits dug below the water table, or from monitoring wells. Rain water was sampled by the WSC at their office in Fort Simpson using a bulk rain gauge. Rain samples were transferred to sample bottles on the same day or the next day following each rain event to minimize the effects of evaporation. Depth-integrated samples of the snowpack were collected during the annual snow survey in March. In May 2002, snow samples were also collected near Main Fen prior to the onset of spring runoff. The oxygen and hydrogen isotopes were analyzed by the standard CO_2 equilibration (Epstein and Mayeda, 1953) and chromium reduction (Gehre et al., 1996) techniques, respectively. The isotopic composition was expressed as δ value with respect to the Vienna-Standard Mean Ocean Water (Gonfiantini, 1981). The concentration of major ions was analyzed using atomic absorption spectroscopy and ion chromatography. The samples for chemical analysis were filtered using 0.45-μm membrane filters. The EC of water samples was measured in a laboratory at 25 °C. The EC of some samples was also measured in the field at various temperatures. The field-measured EC values were converted to the values corresponding to 25 °C assuming the linear temperature–EC relationship (Hayashi, 2004).

4. Results

4.1. Isotopic composition of precipitation

The local meteoric line (LMWL) is defined for the $\delta^{18}O$ and δ^2H of individual precipitation samples collected during March–August 1999 and June–September 2000 using the least-squares regression (Fig. 4). The linear correlation is described by

$$\delta^2H = 7.6\delta^{18}O - 2.0 \tag{1}$$

which is similar to the LMWL determined by Gibson et al. (1993a) at Manners Creek located 14-km southeast of Fort Simpson (Fig. 4). The isotopic composition of individual precipitation samples had a large seasonal variability. The $\delta^{18}O$ ranged between -34.7 and $-24.6‰$ for March–April samples, between -28.3 and $-14.2‰$ for May–June, and between -21.7 and $-12.2‰$ for July–September. Volume-weighted mean $\delta^{18}O$ was $-19.9‰$ and δ^2H was $-150‰$ for all rain samples collected during April–June 1999 and July–September 2000.

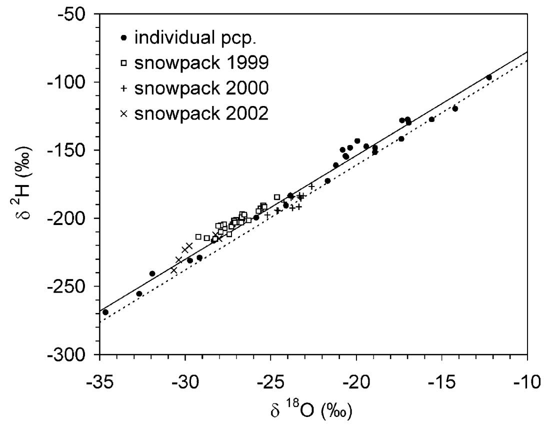

Fig. 4. Isotopic compositions of precipitation and depth-integrated snowpack samples. The solid LMWL is based on the precipitation samples. The dashed LMWL is by Gibson et al. (1993a).

The snow samples in 2002 had an average $\delta^{18}O$ of $-29.5‰$ and δ^2H of $-223‰$ (Fig. 4), which were close to -29.6 and $-228‰$, respectively, measured in 1990 by Gibson et al. (1993b). However, the snowpack had considerably heavier compositions with average $\delta^{18}O$ and δ^2H of -27.0 and $-202‰$, respectively, in 1999 and -23.8 and $-187‰$ in 2000 (Fig. 4). In 1999 the daily mean temperature in Fort Simpson reached 5.1 °C on March 23 and remained above 0 °C on March 20–25 (MSC, 2002a) indicating that some melting and refreezing of the snowpack occurred before the snow sampling on March 23–30. Similarly, the daily mean temperature in 2000 reached 4.6 °C on March 27 before the snow sampling on March 30. This may have resulted in the isotopic fractionation (Cooper, 1998) because some early melt water can infiltrate into the dry surface layer resulting from desiccation (Bowling et al., 2003). In 2002, on the other hand, the snowpack samples collected on May 5 (not shown in Fig. 4), before the major snowmelt runoff started, had an almost identical isotopic composition as the March 2002 samples.

Table 1
Fort Simpson annual and summer (May–September) precipitation for each hydrologic year (October 1–September 30), average snow water equivalent (SWE) in late March from snow survey data, and total annual runoff of Scotty Creek, all reported in mm

	Normal	1999	2000	2001	2002
Period		10/98–09/99	10/99–09/00	10/00–09/01	10/01–09/02
Total pcp.	369	409	431	431	412
May–Sep. pcp.	221	238	296	316	269
SWE	n/a	90	101	n/a	142
Total runoff	n/a	96	139	161	197

n/a indicates the data not available. Normal precipitation is for 1971–2000, and total runoff includes winter runoff.

This may indicate that the isotopic fractionation during the ripening of snowpack was negligible. Further study will be required to evaluate the isotopic fractionation prior to the snowmelt runoff.

4.2. Water balance and chloride mass balance

The average snow water equivalent (SWE, mm) was similar between 1999 and 2000, but was much higher in 2002 (Table 1). As a result, the peak discharge at the outlet of Scotty Creek during the snowmelt in 2002 was much higher than in other years (Fig. 5). The peak discharge in 2002 was the highest recorded since the gauging station was installed in 1994, and the discharge data from the neighbouring Jean-Marie River indicate that the peak discharge in 2002 was the highest in the 30-year record starting in 1973 (Johnson, 2002; WSC, unpublished data). Total discharge in 1999 was 1.46×10^7 m^3, or equivalent to 96 mm of annual runoff from the 152-km^2 basin (Table 1). The hydrograph in 1999 was dominated by snowmelt as summer precipitation was only slightly above the 1971–2000 normal (Table 1). Summer precipitation in 2000–2002 greatly exceeded the normal and, as a result, considerable runoff was generated by summer storms (Fig. 5).

Over the four-year period 1999–2002, while the cumulative precipitation was 1683 mm, only 593 mm discharged from Scotty Creek (Table 1). Assuming that the difference was lost to evapotranspiration, the average annual evapotranspiration of this four-year period was 273 mm/yr. Claassen and Halm (1996) showed that a chloride mass balance can be used to estimate the basin-scale evapotranspiration when the lithologic source of chloride is negligible. The thick organic soils covering the Scotty Creek basin contain little chloride. The underlying mineral soil is mainly derived from clay-rich glacial till having low hydraulic conductivity. The extensive literature on the hydrogeology of clay-rich glacial till in western Canada suggests that active flow of groundwater is limited to a relatively shallow (< 10 m) local system (Hayashi et al., 1998a) and that pre-Holocene chloride within the active flow system has been flushed out (Hayashi et al., 1998b). Therefore, Scotty Creek provides favourable conditions for applying the chloride method, where precipitation is the dominant input and stream flow is the dominant output of chloride. Chloride within the basin drainage system takes multiple transport pathways including overland flow, subsurface flow through the unfrozen peat layer, and groundwater flow through the mineral soil; and eventually enters the main channel of Scotty Creek and reaches the outlet.

Using the chloride method, evapotranspiration E_t is given by

$$E_t = P(C_s - C_p)/C_s \tag{2}$$

where P is the annual precipitation, and C_s and C_p are the volume-weighted average chloride concentration in stream water and precipitation, respectively.

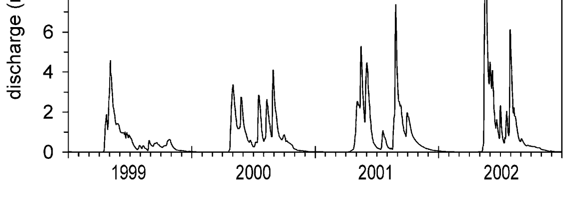

Fig. 5. Daily average discharge of Scotty Creek at the outlet.

Forty-three water samples were collected at the outlet between March and December during 1999–2002 and analyzed for chloride. The volume-weighted average concentration was calculated by summing the product of the chloride concentration and the stream discharge at the time of sample collection, and dividing the total by the sum of all discharge values. The average C_s for the four-year period was 0.151 mg/l. The average C_p ($=0.044$ mg/l) is given by the 10-year mean (1992–2001) of chloride in precipitation reported in the NatChem database (MSC, 2002b) at Snare Rapids, located 400-km northeast of Scotty Creek. This value is similar to the NatChem data from other stations in the interior western Canada (0.04 mg/l) presented by Hayashi et al. (1998b). The average precipitation for the hydrological years 1999–2002 was 421 mm/yr (Table 1). Therefore, Eq. (2) gives $E_t = 298$ mm/yr, which agrees with the hydrometric estimate of 273 mm/yr. A simple arithmetic average concentration of the 43 samples was 0.133 mg/l. Using this value for C_s in Eq. (2) gives $E_t = 282$ mm/yr, which is also in agreement with the hydrometric estimate. These results suggest that the chloride method has a great potential as a tool for estimating the basin-scale evapotranspiration in ungauged basins. Chloride concentration in Scotty Creek had a large seasonal variation ranging from less than 0.01 to 0.37 mg/l, presumably reflecting evaporative enrichment in summer and changes in the relative contribution of various source waters. Therefore, the average of numerous samples collected at various times was necessary to estimate a representative value for C_s. Using a few samples of base flow as a surrogate, as suggested by Claassen and Halm (1996), would have given an inaccurate result because the concentration varied substantially even during the low-flow period. Successful application of the chloride method to ungauged streams in northern Canada will require developing a reliable method to estimate the average concentration from a limited number of samples.

4.3. Chemical and isotopic composition surface and subsurface water

Goose Lake, the headwater of Scotty Creek, had a considerably enriched isotopic composition (Fig. 6) indicating that evaporation was the dominant output from the lake (Gibson et al., 1993a). Surface water samples from the study nodes along Scotty Creek had the compositions ranging between those of Goose Lake and soil water collected from the active layer of peat plateau (Fig. 6). The isotopic composition of the water sampled at the outlet was generally distributed along the LMWL (Fig. 6).

The δ^{18}O during a high-flow period (June 10, 2000) was highest at Goose Lake and generally decreased downstream (Fig. 7a). The δ^{18}O of the sample collected at the North Arm node on the same

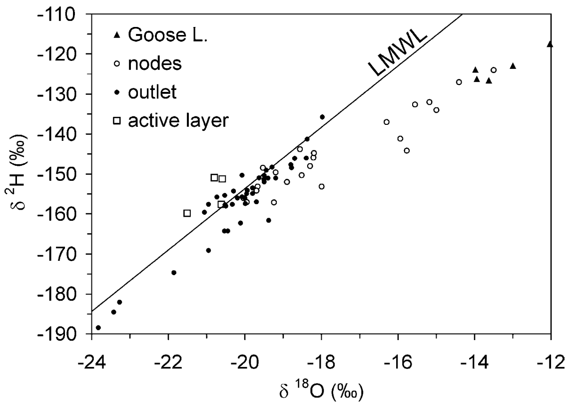

Fig. 6. The isotopic composition of surface water sampled at Goose Lake, the study nodes along the Scotty Creek drainage system, and the outlet. Also shown is the isotopic composition of subsurface water from the active layer of the peat plateau adjacent to Main Fen, and the local meteoric water line (LMWL).

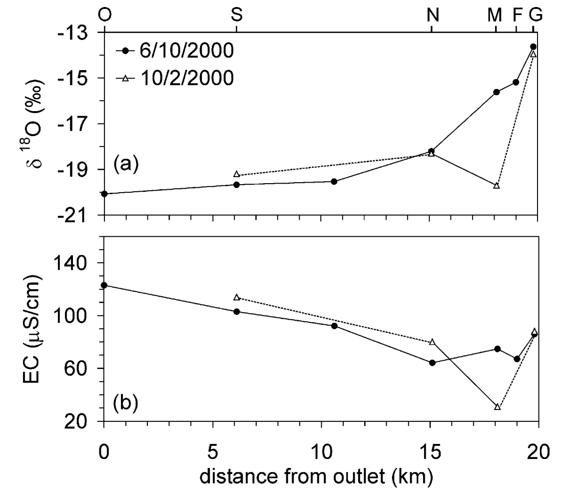

Fig. 7. The δ^{18}O (a) and EC (b) of stream water in relation to the distance from the outlet. The location of study nodes are indicated on the top axis; Goose Lake (G), First Lake (F), Main Fen (M), Next Lake (N), South Arm (S), and outlet (O).

date was $-20.0‰$ (not shown in Fig. 7), not appreciably different from the South Arm sample. A similar pattern was also observed for δ^2H (not presented). The $\delta^{18}O$ at Main Fen became substantially lower during a low-flow period (October 2, 2000), but the values stayed fairly constant at other nodes. A similar pattern was also observed in 1999 (not shown). The EC of stream water gradually increased downstream during the high-flow period in 2000 (Fig. 7b). The EC of the North Arm sample on the same date was 141 μS/cm, much higher than that of the South Arm sample (103 μS/cm). The EC at the outlet (123 μS/cm) was approximately equal to the average of the two tributaries. This is consistent with the discharge measurement of Quinton et al. (2003), who reported that the contribution from the two tributaries were approximately equal to June 10, 2000. The EC at Main Fen became considerably lower during the low-flow period (Fig. 7b), while the EC at other nodes stayed fairly constant.

The EC was most strongly correlated with the concentrations of Ca, Mg, and HCO_3 as indicated by a linear relationship (Fig. 8) between Ca + Mg and EC for water samples collected in June 1999. These ions are generally derived from the dissolution of carbonate minerals. A groundwater sample from the piezometer screened in the mineral soil under Main Fen had an EC of 1290 μS/cm and Ca + Mg of 13.9 mequiv./l. The EC to Ca + Mg ratio of this sample is very close to those of the stream water samples plotted in Fig. 8, suggesting that the source of Ca and Mg in stream water is mineral-rich groundwater. However, the anion composition of the groundwater sample was dominated by SO_4 (10.7 mequiv./l) compared to HCO_3 (5.3 mequiv./l). This was markedly different from stream water samples, in which HCO_3 represented more than 98% of negative charge. The hydraulic conductivity of mineral soil estimated by a slug test (Freeze and Cherry, 1979, p. 341) on the piezometer was 5×10^{-8} m/s, and the magnitude of vertical hydraulic gradient between the piezometer screen and the surface water was less than 0.005, indicating a small flow velocity in the order of 8 mm/yr or less. Therefore, the groundwater discharging through the peat layer above the mineral soil has a large residence time for complex geochemical processes that reduce SO_4 and generate HCO_3 (Spratt and Morgan, 1990).

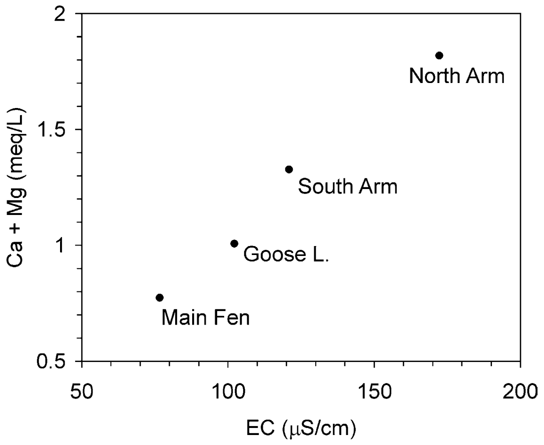

Fig. 8. Correlation between electrical conductivity (EC) and the concentration of Ca and Mg in June 1999.

A detailed discussion of geochemical processes is beyond the scope of this paper, but in general the high contrast in EC between surface water and mineral-rich groundwater suggests that a relatively small addition of groundwater can account for the gradual downstream increase of EC. In contrast to the dramatic difference in EC, the isotopic composition of the groundwater sample ($\delta^{18}O = -16.7‰$ and $\delta^2H = -150‰$) was similar to the average of six samples (-16.6 and $-141‰$) collected from the fen during 1999–2002. Therefore, the addition of mineral-rich groundwater to surface water is expected to have minor effects on the isotopic composition of stream water.

The subsurface water sampled from the active layer of the peat plateaus had low EC (20–40 μS/cm) and depleted isotopic composition (Fig. 6). The isotopic composition of the active layer waters was similar to the average composition of summer rain ($\delta^{18}O = -19.9‰$ and $\delta^2H = -150‰$), but slightly lighter indicating the influence of winter precipitation. Subsurface runoff from the active layer is expected to be an important source of water for Scotty Creek during the snow-free period. The top of the frozen soil (i.e. the frost table) represents the lower boundary of the thawed, saturated zone in this active layer (Fig. 3). A rainfall-induced water table rise on the peat plateaus dramatically increases subsurface drainage toward the fen, since not only is the hydraulic gradient toward the fen increased, but the permeability increases by 2–3 orders of magnitude between

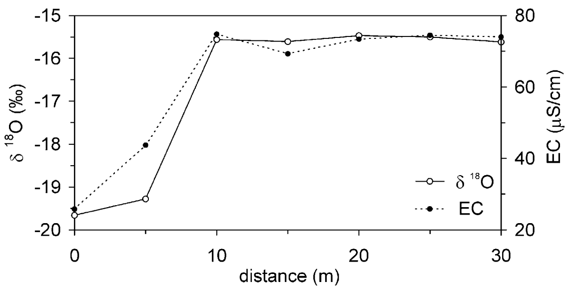

Fig. 9. Variation of δ^{18}O and electrical conductivity (EC) in Main Fen on June 20, 2000 plotted against the distance from the edge of the fen.

the bottom and top of the active layer (Quinton and Gray, 2003).

Fig. 9 plots the δ^{18}O and EC of surface water at Main Fen on June 10, 2000 in relation to the distance from the edge of the fen. The δ^{18}O and EC values near the edge (< 10 m), immediately below the peat plateau, were very similar to those of the subsurface water in the active layer (Fig. 6), indicating the influence of subsurface drainage from the peat plateau (Fig. 3). In the central part of the fen, the δ^{18}O and EC values were generally constant (Fig. 9) and was similar to the values measured at First Lake on the same day (Fig. 7). It is interesting to note that the δ^{18}O and EC in Main Fen were similar to the values at First Lake during the high-flow period (Fig. 7, June 10), but dropped to the values similar to those of the active layer of the peat plateaus during the low-flow period (Fig. 7, October 2). The dramatic seasonal shift in Fig. 7 suggests that the drainage from the peat plateaus supplied the majority of water to Main Fen in the fall. This may imply that Main Fen had become hydrologically disconnected from First Lake. In contrast to Main Fen; Goose Lake, First Lake, and Next Lake did not show a dramatic compositional change (Fig. 7) presumably because of their large volume to buffer short-term fluctuations in input water compositions.

Water samples from flat bogs located to the east of Main Fen (Fig. 2) had low pH (4.6–5.5) and EC (30–50 μS/cm) values, typical of ombrotrophic peat bogs. Their low EC values clearly indicate that they are not 'flow through' features like channel fens. Field observations at these flat bogs confirmed that they receive subsurface drainage from peat plateaus. It was also noted from observations on the ground and from the air during helicopter reconnaissance that surface and shallow subsurface water drains from these flat bogs to the adjacent channel fen. However, there are many other flat bogs that are completely encircled by peat plateaus, and therefore appear to be internally drained (Fig. 2).

4.4. Time series of stream water

Fig. 10a shows the seasonal variation of δ^{18}O at the outlet of Scotty Creek. The δ^{18}O was the lowest immediately after the commencement of snowmelt runoff, indicating a strong influence of the lighter snowmelt water (Fig. 4). Snowmelt runoff started about April 20 in 1999–2001 (Fig. 5), but did not start until May 11 in 2002. The δ^{18}O value gradually increased in summer and reached a stable value of -18.5 to -19.5‰ by August. The δ^{2}H also had the similar variation (not presented) and reached a stable summer value of -142 to -150‰. This presumably represents the mixing of the enriched Goose Lake water (Fig. 6), summer precipitation, and the melt water from the active layer (Fig. 6); as well as evaporative enrichment in fens and lakes. The δ^{18}O value kept decreasing after the freeze up, which may indicate the depletion of heavier isotope species in the liquid phase by freezing (Gibson and Prowse, 1999).

Fig. 10b shows the seasonal variation of EC at the Scotty Creek outlet. The EC was 50–60 μS/cm during the peak of the snowmelt period and gradually increased until it reached a stable value of 170–180 μS/cm in July–August. The lower value during the snowmelt period indicates a higher contribution of fresh snowmelt water having very

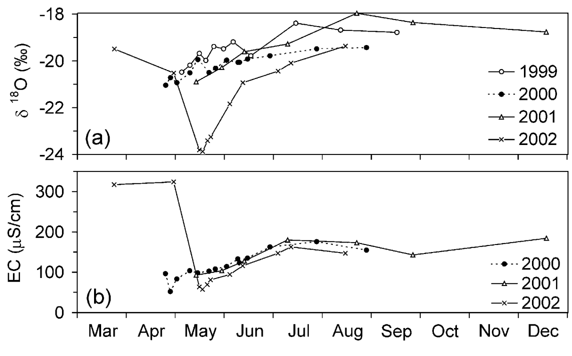

Fig. 10. Seasonal variation of δ^{18}O (a) and electrical conductivity (b) at the outlet of Scotty Creek.

little solutes. The EC continued to increase in winter when the flow in Scotty Creek was very small, and reached above 300 μS/cm in March (Fig. 10b). This is probably due to the exclusion of ions from the ice forming in lakes and channel fens and the increased contribution of mineral-rich groundwater. The values and patterns of EC were fairly similar between years except that the timing of snowmelt runoff was delayed in 2002. In contrast to those of δ^{18}O varied from year to year, indicating the inter-annual variability in the isotopic composition of source waters, most notably the snowpack (Fig. 4).

4.5. Snowmelt hydrograph separation

The time series of tracer concentrations (Fig. 10) can be used to separate the contribution of event and pre-event waters in stream hydrographs, when the compositions of source waters are known (Sklash and Farvolden, 1979). In cold regions underlain by frozen soil, isotope tracers have been used to separate the contribution of snowmelt (event) water from that of non-snowmelt water, where in-channel storage was assumed negligible (Cooper et al., 1993; McNamara et al., 1997). In contrast the present study tries to evaluate the storage of runoff water in the network of channel fens and lakes, which is one of the most critical processes in the water cycle.

The rise of stream discharge at the outlet of Scotty Creek generally occurs following a period of warm weather. For example, in 2000 the daily mean air temperature at Fort Simpson (MSC, 2002a) was above 0 °C for 13 consecutive days preceding April 20, when the discharge started increasing. The peat plateaus remain frozen, except for the top 5–10 cm, during the entire melt period as was confirmed by the soil temperature data at the peat plateau adjacent to Main Fen in 2000 and 2002. The surfaces of wetlands had much thinner snow cover than peat plateaus (Onclin et al., 2000; Onclin and Best, 2002). The runoff from peat plateaus drains onto the snow-free, frozen surface of wetlands and lowers the albedo, thereby enhancing the ablation of ice and underlying frozen peat. No data on the ice ablation rate was collected in this study, but Woo and Heron (1987) reported 50 cm of ice ablation in one week at a subarctic wetland-river complex in northern Ontario, Canada. Therefore, it is reasonable to expect that the water stored over winter in wetlands and lakes starts to mix with snowmelt water shortly after the commencement of runoff.

Snowmelt runoff at Scotty Creek occurs when the soil is still frozen and relatively impermeable, suggesting that the contribution of frozen soil water to runoff is minor. A large amount of surface runoff flows through the drainage system over a few weeks so that the contribution of regional groundwater discharge through relatively impermeable mineral soil is expected to be negligible. In contrast, a large amount of water stored in lakes and wetlands is expected to have a major influence on the composition of stream water because these lakes and wetlands are part of the drainage system. Therefore, the following analysis assumes that the spring runoff in Scotty Creek comes from only two sources, the snowpack of the current year and the water stored over winter in lakes, stream channels, and wetlands. The latter represents mixed inputs of various source waters during the unfrozen period, such as direct precipitation, subsurface runoff from the active layer, and groundwater discharge through the mineral soil. The period of analysis is selected so that the contribution of rain is minor compared to other sources.

In the two-component system, the fraction of event water (x) is given by (Rodhe, 1998)

$$x = (C_s - C_{pe})/(C_p - C_{pe}) \tag{3}$$

where C_s, C_e, and C_{pe} are tracer concentrations in stream water, event water (snowpack), and pre-event water (over-winter storage), respectively. Three tracers (EC, δ^{18}O, and δ^2H) were used with Eq. (3). The isotopic composition of the snowpack showed a large degree of inter-annual variability (Fig. 4). It is not certain if this variability reflects the variability of snow itself or the isotopic fractionation in the snowpack. Nevertheless, the mean δ^{18}O and δ^2H of all snowpack samples for each year (Fig. 4) were used as C_e. Measured EC values of snowpack samples were all less than 10 μS/cm and, hence, $C_e = 5$ μS/cm was assumed.

It is not easy to define a single value of C_{pe} to represent the average composition of water stored over winter in numerous channel fens and lakes. The EC and δ-values of the samples collected at the outlet was used as a surrogate of C_{pe}. These values presumably represent the average composition of all water sources weighted by the amount of input from each source during the ice-free, base-flow period. Note that the amount of input is not necessarily proportional to the amount of stored water, and the relative contribution of different sources may change between low-flow and high-flow period. Two values of C_{pe} were used to reflect uncertainty in this parameter. For EC, the first value is 160 μS/cm based on the samples taken during base-flow periods in July and August (Fig. 10b). This value is most likely biased by the contribution of mineral-rich groundwater discharging at the lower portion of the basin because most water samples from the upper portion had much lower EC throughout the frost-free season (Fig. 7b). The second value 130 μS/cm was used to represent the average concentration in mid June, shortly after the snowmelt runoff (Fig. 7b). The actual value of C_{pe} is most likely between the two extremes. The uncertainty in C_{pe} for δ^{18}O was smaller than for EC because δ^{18}O values were relatively uniform downstream of Next Lake (Fig. 7a), which presumably served as a buffer. The late summer value of −18.5 to −19.5‰ was used as an estimate of C_{pe}. Similarly, the late summer value of −142 to −150‰ was used as an estimate of C_{pe} for δ^{2}H.

The cross-plot of multiple tracers, or mixing diagram, is a useful tool for examining the appropriateness of end member values (Brown et al., 1999). Fig. 11 shows the assumed ranges of C_e and C_{pe} of the three tracers, where error bars for the snowpack δ^{18}O and δ^{2}H indicate the 95% confidence limits. Note that the error bars are larger for the 2002 snowpack samples having smaller sample size ($n = 6$) compared to the 2000 snowpack samples ($n = 14$). The water samples collected during snowmelt runoff in 2000 and 2002 show a linear trend between the pre-event and event composition (Fig. 11). One exception is the sample taken on April 28, 2000 (marked with an arrow in Fig. 11a), which had a substantially lower EC than expected. The reason for the deviation is uncertain. Despite this particular sample, it is reasonable to assume that the composition of stream water during snowmelt runoff can be represented by linear mixing of the two end members.

Eq. (3) was applied to the EC data of 2000 (Fig. 13b) using $C_{pe} = 130$ μS/cm (bottom curve separating event and pre-event water) and 160 μS/cm (top curve) representing a range of uncertainty. The amount of precipitation was relatively small

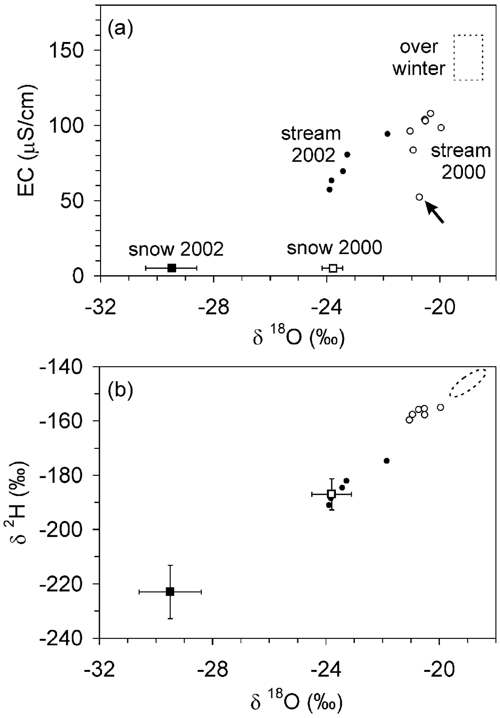

Fig. 11. Mixing diagrams for δ^{18}O and electrical conductivity (a) and δ^{18}O and δ^{2}H (b). Legends are shown in (a). Error bars indicate 95% confidence limits. The arrow in (a) indicates the sample collected on April 28, 2000.

(Fig. 12a), during the period of analysis (April 19–May 23). The stream discharge in Fig. 12b is reported as daily runoff, which is the total daily discharge divided by the basin area. The total runoff during April 19–May 23 was 33 mm. The estimated contribution of the event water was 9.8 mm for $C_{pe} = 130$ μS/cm and 14 mm for $C_{pe} = 160$ μS/cm. Therefore, the relative contribution of event water was between 31 and 44% (Table 2). For the analysis of δ^{18}O data (Fig. 12c), the bottom curve was calculated using $C_{pe} = -19.5$‰ and $C_e = -24.2$‰, and the bottom curve was calculated using $C_{pe} = -18.5$‰ and $C_e = -23.4$‰. The estimated relative contribution of event water was between 25 and 44% (Table 2). A similar analysis of δ^{2}H data (not presented) gave the relative contribution between 17 and 35% (Table 2). Eq. (3) was also used to analyze the snowmelt runoff in 2002 (Fig. 13). The amount of precipitation was relatively small (Fig. 13a) during the period of analysis (May 11–June 5), except for a snow event on May 27–28. The total runoff

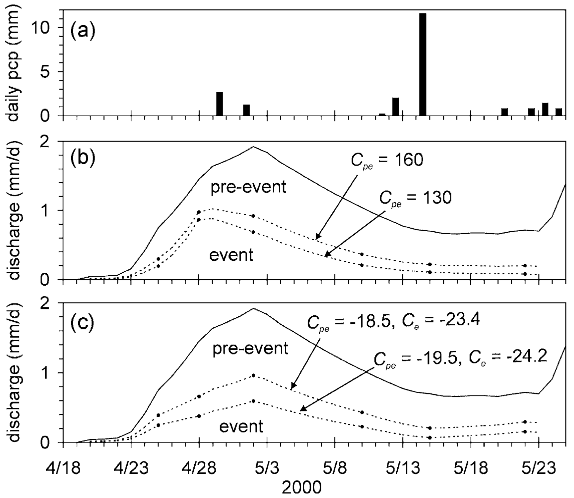

Fig. 12. Daily precipitation at Fort Simpson for 2000 (a) and hydrograph separation using electrical conductivity (b) and δ^{18}O (c). Solid lines indicate total discharge. Solid circles indicate the contribution of event water based on the water sample and dashed lines indicate interpolated results.

during this period was 86 mm. The estimated contribution of event water in 2002 was higher than that of 2000 (Table 2), most likely because the SWE was much greater (Table 1) and the snowmelt happened much more rapidly.

5. Discussion

5.1. Processes on the peat plateau affecting the composition of snowmelt runoff

Hydrograph separations using EC and stable isotopes all gave qualitatively similar results, indicating a major contribution of pre-event water, but estimated values had a large degree of variability (Table 2). The varying results suggest that unaccounted processes or sources may have affected the chemical and isotope compositions of stream water. The isotopic composition of melt water from the peat plateau may be altered due to isotopic fractionation. Cooper (1998) and Rodhe (1998), observing that the snowmelt composition changed during the melt period, suggested that the direct sampling of snowmelt water was necessary to estimate the event-water concentration accurately. The early snowmelt water, having a potentially lighter isotopic composition (Cooper, 1998), may infiltrate into the still frozen, but partially saturated peat near the surface, thereby increasing the ice content and making it impermeable to the later snowmelt water having a heavier isotopic composition (Slaughter and Kane, 1979; Roulet and Woo, 1986). Alternatively, when the melt water mixes with the soil water in the thawed zone, the isotopic composition of melt water

Table 2
Contribution (%) of the event (snowmelt) water to snowmelt runoff estimated by the EC, ^{18}O, and ^{2}H methods

Tracer	2000	2002
EC	31–44	44–54
^{18}O	25–44	34–46
^{2}H	17–35	41–56

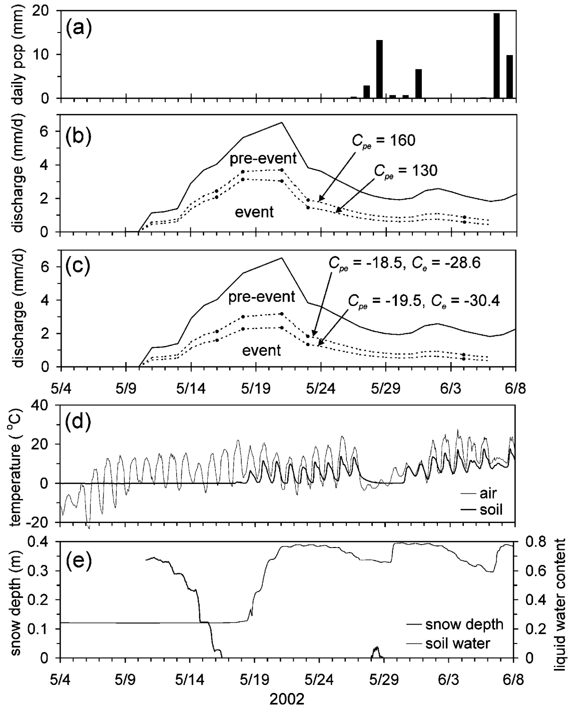

Fig. 13. Daily precipitation at Fort Simpson for 2002 (a), hydrograph separation using electrical conductivity (b) and δ^{18}O (c). Air temperature and soil temperature at 5 cm (d), and snow depth and liquid water content at 10 cm at the peat plateau of the Main Fen site (e). Legends for (b) and (c) are the same as Fig. 12. Snow-depth data before May 10 were unavailable due to equipment failure.

becomes heavier, because the water in the active layer has a composition similar to the pre-winter storage in wetlands. Gibson et al. (1993b) noted a 4–5‰ higher $\delta^{18}O$ values in snowmelt runoff in pipes and rills, compared to snowpack, and attributed it to mixing of the pre-event subsurface water in micro-depressions on permafrost slopes.

Detailed hydrological data were collected at the Main Fen site in 2002 to improve our understanding of these processes. The site is located on a peat plateau typical of this region having a very low topographic relief. Therefore, the data represent the hydrometeorological condition of the basin reasonably well. Snowmelt probably started on May 6 as indicated by air temperature data (Fig. 13d), but the stream discharge at the outlet did not start rising until May 11 (Fig. 13b). The soil temperature at 5 cm stayed at the freezing point until the snowpack disappeared on May 16 (Fig. 13e). Liquid water content at 10 cm remained at the residual value under the frozen condition (Spaans and Baker, 1996) until May 18, indicating that the infiltration of melt water did not reach this depth until the soil above started thawing. These findings suggest that the interaction of snowmelt water with peat on the peat plateau is limited to the top 5–10 cm. Therefore, the processes affecting the chemical and isotopic composition of snowmelt runoff probably occur within the snowpack and/or a very thin layer of the surface soil. Improving the accuracy of hydrograph separation will require a further study of isotopic and chemical processes within the snowpack and over-winter redistribution of water and solutes in the soil associated with freeze–thaw cycles.

5.2. Storage and mixing of water in wetlands and lakes

Bowling et al. (2003) estimated the storage of snowmelt runoff in a wetland-dominated, 471-km^2 basin on the Alaskan North Slope, where the period of snowmelt runoff was clearly defined. It is difficult to perform a similar water-balance analysis for Scotty Creek because the end of snowmelt period is obscured by summer storms (Fig. 5). Nevertheless, if we assume the snowmelt runoff of 33 mm for 2000 and 86 mm for 2002, the available snow water equivalent on the ground (Table 1) minus snowmelt runoff is 68 mm in 2000 and 56 mm in 2002. These numbers represent approximate amounts of snowmelt water stored in the basin. The stored water is eventually released to evapotranspiration and drainage during summer months (Bowling et al., 2003).

Mixing ratios presented in Table 2 give an indication of the total amount of pre-event water stored in the basin that is readily mixed with snowmelt water

$$S_r = \mathrm{SWE}(1 - x)/x \tag{4}$$

where S_r (mm) is the amount of pre-event water stored over winter, SWE is the snow water equivalent prior to melt, and x is event water fraction (Eq. (3)). Eqs. (3) and (4) are based on the assumption that the water represented by S_r and SWE become completely mixed, and a portion of the mixture discharges at the outlet while the rest is retained in a ‘reservoir’. Assuming $x = 0.3$–0.4 for 2000 (Table 2) and noting that SWE = 101 mm (Table 1), S_r is estimated to be 152–236 mm. Similarly, using $x = 0.4$–0.5 and SWE = 147 mm for 2002 in Eq. (4) gives $S_r =$ 142–213 mm. The average basin discharge for a period 1999–2002 was 149 mm/yr (Table 1). Therefore, S_r is comparable to or greater than annual discharge. A large part of S_r probably consists of surface water and ice in lakes and wetlands, and subsurface ice in the highly permeable top layer of organic soils under wetlands.

Quinton et al. (2003) estimated that approximately 3.5% of the basin is covered by Goose Lake and other smaller lakes. No depth data exist for lakes in the basin, but Gibson and Prowse (1999, Fig. 5) measured the $\delta^{18}O$ profile of the ice on Goose Lake in March 1997. Assuming that the lake was essentially a closed system, the fraction of liquid water remaining in the lake can be estimated from the equation of Rayleigh-type fractional crystallization process (Gibson and Prowse, 1999, Eq. (4)). Using this method with an isotopic separation factor ($\varepsilon_{\text{ice-water}}$) of 2.94‰ for $\delta^{18}O$, it was estimated that approximately 25% of water remained as liquid, when the ice thickness was 75 cm. Assuming an ice density of 0.9, the total depth of water before freezing is 97.5 cm. Using 100 cm as an estimate of average depth of all lakes in the basin, the lakes covering 3.5% of the basin is equivalent to 35 mm of water distributed over the basin, which is much smaller than the estimates of S_r above. Therefore, the majority of S_r is probably stored in wetlands (i.e. fens and bogs) covering 30% of the basin (Quinton et al., 2003). Wetlands are distributed throughout the basin and become connected to and disconnected from the main drainage system depending on the water level, as indicated by the dramatic change in chemical and isotopic compositions at Main Fen (Fig. 7). The analysis presented in this study assumes that the stream samples at the outlet represented the average composition of water stored in all wetlands and lakes that are part of the drainage system. Successful application of the chloride mass balance suggests that this assumption is reasonable for the steady-state analysis over multiple years. For the transient analysis of the snowmelt event, however, the uncertainty in the composition of pre-event water (Fig. 11) represents a major source of error.

6. Conclusions

The Scotty Creek basin consists of three major land-cover types (peat plateaus, flat bogs, and channel fens) having different hydrologic functions. Peat plateaus are underlain by permafrost, which prevents the vertical flow of water. The lateral transfer of water from peat plateaus to flat bogs and channel fens normally occurs as subsurface runoff through the active layer. The water in the active layer of peat plateaus had low EC, 20–40 μS/cm, indicating little contact with mineral soils. Flat bogs generally have the water table below the ground and receive direct precipitation and the runoff from peat plateaus. Some flat bogs drain water to channels fens, while others appear to be internally drained. Water in flat bogs had low pH (4.6–5.5) and EC (30–50 μS/cm) typical of ombrotrophic peat bogs despite receiving water inputs from peat plateaus. Channel fens are flow-through features forming a network that is part of the basin drainage system. The EC and isotopic composition of water in a channel fen during high-flow periods were similar to those of upstream lakes owing to its flow-through nature. However, during low-flow periods, the EC and isotopic composition shifted to values similar to those of the active layer on the surrounding peat plateaus, implying that this channel fen was hydrologically disconnected from the basin drainage system.

The chloride balance method gave 280–300 mm/yr as average evapotranspiration over the entire basin, consistent with the hydrometric estimate (precipitation minus runoff) of 275 mm/yr indicating an excellent potential of this method to ungauged basins, where hydrometric data are not available. However, the measured chloride concentration had a large degree of variability, suggesting that a sufficiently large number of samples are necessary to estimate an average concentration.

EC, oxygen-18, and deuterium were used to separate hydrographs for the direct contribution of snowmelt (event water) and the water stored in the basin over winter (pre-event water). All three tracers indicated that the direct snowmelt contribution was less than half of total discharge, indicating an importance of the water stored over winter in the interconnected lakes and channel fens. However, discrepancies between different tracers suggest the limitation of applying conceptually simplified two-component chemical and isotope hydrograph separations to a complex system. The total amount of pre-event water stored over winter in the basin was estimated to be 140–240 mm. This magnitude is comparable to the average annual basin discharge (149 mm/yr), suggesting the importance of considering the wetland storage in the hydrological models of the Scotty Creek basin, which also represents numerous other river basins in the wetland-dominated, discontinuous permafrost region.

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