Biogeochemistry **47:** 239–267, 1999.

The effect of permafrost on stream biogeochemistry: A case study of two streams in the Alaskan (U.S.A.) taiga

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Received 11 November 1998; accepted 10 December 1998

Key words: Alaska, biogeochemistry, permafrost, streams, taiga, watershed

Abstract. Understanding interactions between permanently frozen soils and stream chemistry is important in predicting the effects of management, natural disturbance and changing permafrost distribution on stream ecosystems and nutrient budgets in subarctic watersheds. Chemical measurements of groundwater, soil water and stream water were made in two watersheds in the taiga of interior Alaska. One watershed (HiP) had extensive permafrost and the other (LoP) had limited permafrost. Soil water collected within the rooting zone (0.3–0.5 m) in both watersheds was high in dissolved organic carbon (DOC), dissolved organic nitrogen (DON) and dissolved inorganic nitrogen (DIN) but low in dissolved minerals (dominantly Ca, Mg and Na) and conductivity. The reverse was true for groundwater from springs and wells. Permafrost in the HiP basin prevented deep percolation of water and generated stormflows rich in DOC. The presence of permafrost in HiP resulted in higher fluxes of DOC, DON and DIN into stream water from upland soils.

Introduction

The washing of dissolved materials from terrestrial ecosystems into streams represents an important component of the biogeochemical balance of many ecosystems and it can have a significant influence on the ecology of stream organisms. Understanding the scale of these fluxes and the processes controlling them is important in developing accurate biogeochemical models and in facilitating effective watershed management.

There is little published material about the processes and scales of dissolved material transport through high latitude watersheds with permafrost. Studies for various ecosystems reviewed by Likens and Bormann (1995) show a preponderance of data from temperate or tropical systems. Chemical sources and fluxes have been studied in boreal areas below the southern limit of permafrost (Wallis et al. 1981; Naiman 1982; Lock & Ford 1986) and in Arctic watersheds completely underlain by permafrost (Peterson et al. 1986; Lock et al. 1989; Shaver et al. 1991; Peterson et al. 1992; Kling et al. 1992; Everett et al. 1996; Oswood et al. 1996; McNamara et al. 1997). However, there is very little published information on watershed biogeochemistry from areas of the Subarctic with discontinuous permafrost. There is a need for more information from the Subarctic because generalizations which hold true for lower latitudes or warmer parts of the Subarctic may not hold true for Subarctic systems with permafrost.

Permafrost has a profound and usually predictable influence on the hydrology of streams in the Subarctic. Streams draining permafrost-dominated watersheds have a more "flashy" hydrology than those draining permafrost-free watersheds (Woo 1986). A "flashy" hydrologic regime is characterized by low baseflows but high stormflows with a rapid onset following snowmelt or rainfall (Ford 1973; Slaughter & Kane 1979; Haugen et al. 1982). The absence of permafrost allows deeper infiltration of precipitation and is thought to allow greater and more sustained baseflows and reduced stormflows (Slaughter & Kane 1979; Woo 1986; Woo & Winter 1993). In the Yukon-Tanana uplands of Interior Alaska the distribution of permafrost is discontinuous and largely a function of aspect. South-facing slopes receive more solar warming and are permafrost-free, whereas north-facing slopes and valley bottoms are often shaded, cooler and underlain by permafrost (Viereck et al. 1983). The balance of warm and cold aspects is, therefore, a controlling factor in the hydrology of individual watersheds throughout this region (Slaughter & Kane 1979; Haugen et al. 1982) and creates a hydrologic mosaic across the northern taiga.

Permafrost reduces dissolved inorganic mineral loads in streams. Ray (1988) compared the stream chemistry of four watersheds (5–11 km^2 in area) with varying amounts of permafrost (approximately 3%–50%) in the Caribou-Poker Creeks Research Watershed (CPCRW) in Interior Alaska. He showed that as estimated permafrost coverage increased, average stream water concentrations of dissolved Ca, Mg, Na, SO_4^{2-} and HCO_3^- decreased. Ice-rich permafrost is relatively impermeable compared to permafrost-free soils (Woo 1986). Ray (1988) proposed that permafrost confined runoff to upper organic soil horizons and reduced transmission through mineral soils thus reducing the dissolution and transport of dissolved inorganic ions. Studies in arctic tundra support this hypothesis. Dissolved mineral concentrations in stream water from a 2 km^2 arctic watershed with 100% permafrost (Everett et al. 1996) are lower than concentrations in stream water from one with 50% permafrost in CPCRW (Ray 1988). Even though permafrost is usually continuous in the Arctic, the dissolved mineral outputs from upland soils increase as the active layer deepens and more water infiltrates mineral soil horizons (Buttle & Fraser 1992; Everett et al. 1996; McNamara et al. 1997).

The objective of this study was to use soil water and groundwater chemistry along with stream water chemistry to refine our understanding of the effects of permafrost on watershed biogeochemistry with a particular focus on DOC (dissolved organic carbon), DON (dissolved organic nitrogen) and DIN (dissolved inorganic nitrogen) fluxes. Using a comparison of two watersheds in CPCRW as a case study, we compared soil water, groundwater and stream water chemistry between a watershed with limited permafrost (LoP, low permafrost) and a watershed with extensive permafrost (HiP, high permafrost). In addition to DOC, DON and DIN, we measured a range of dissolved mineral concentrations (Ca, Mg, Na, Al, Mn, Si and SO_4^2). We used these data to test the following assumptions:

1. Water moving through upper soil horizons is rich in dissolved organic material but low in dissolved inorganic material;
2. Groundwater is low in dissolved organic material but rich in dissolved inorganic material;
3. Shallow runoff makes a much larger contribution than groundwater to total stream discharge in HiP than in LoP.

Materials and methods

Study site

Caribou-Poker Creeks Research Watershed encompasses 10,620 ha and is situated about 50 km northeast of Fairbanks, Alaska (Figure 1). It consists of rolling, forested hills and wide boggy valley bottoms drained by meandering streams with shallow gradients. The watershed (elevation 224 m to 826 m) is representative of much of the Yukon–Tanana Uplands. The region has a cold continental climate. In nearby Fairbanks the average temperature for January (the coldest month) is −24 °C. and that for July (the warmest month) is 17 °C (National Weather Service, unpublished data). Average annual precipitation for the region is 285 mm (National Weather Service, unpublished data) but annual precipitation averages about 500 mm in the upland parts of the watershed (Ford 1973; Ray 1988). Half of all precipitation falls as snow, which commonly covers the ground for an average of 214 days each year (National Weather Service, unpublished data). Streams are normally ice-covered from late October until early May. Because of the high latitude and low sun angles, north aspects receive much less solar warming than south aspects (Koutz & Slaughter 1973). Approximately 30 percent of CPCRW is underlain by

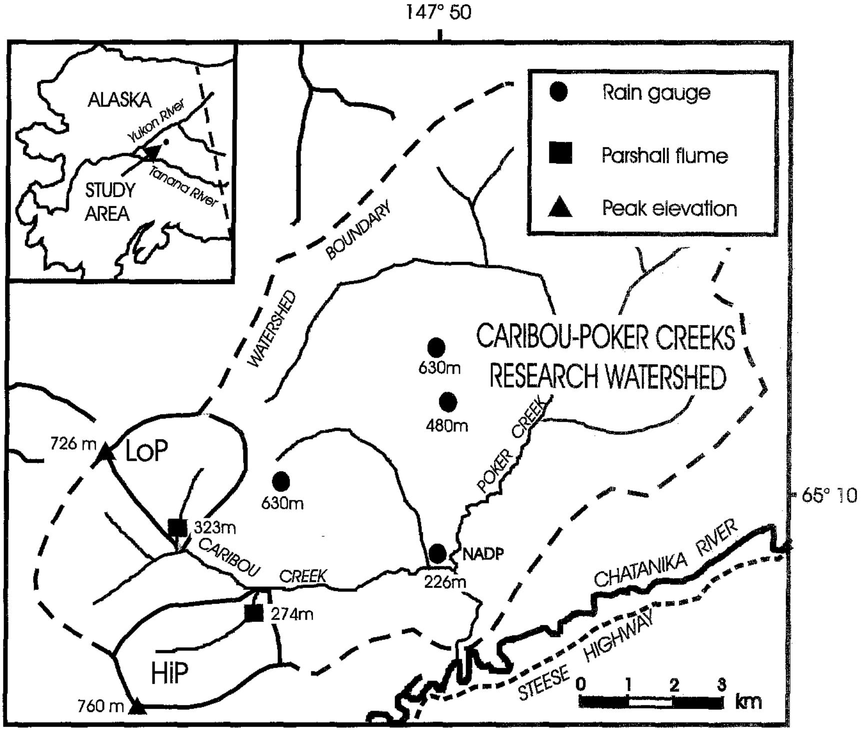

Figure 1. The Caribou-Poker Creeks Research Watershed showing study watersheds LoP (low permafrost) and HiP (high permafrost). Rain gauges and flumes used in this study are also shown. The inset shows the location of the study area relative to the Yukon and Tanana Rivers. Elevations for flumes, rain gauges and peaks are shown in meters a.s.l.

permafrost. This is found mostly on north facing slopes and valley bottoms (Rieger et al. 1972).

LoP and HiP are first-order watersheds (based on US Geological Survey 1:63,360 scale topographic maps) within CPCRW and are generally referenced in watershed literature as C2 and C3, respectively. The terms LoP and HiP have been adopted here for clarity. The two watersheds have similar area (520 and 570 ha, respectively) and relief but differ in their aspect. HiP drains to the north-east and is colder than LoP which drains to the south (Koutz & Slaughter 1973). Estimates of permafrost distribution are based on a soil survey map at 1:30 000 scale with a minimum mapping unit size of 2 hectares (Rieger et al. 1972). Based on this map 3% of the area of LoP and 53% of the area of HiP is underlain by permafrost. Estimates of permafrost extent must be treated as approximate only.

Both study watersheds are underlain by mica schist of the Birch Creek formation covered by a thin loess mantel from the same formation. Valley bottoms consist of deep alluvial silt and gravels over bedrock or loess with permafrost at 1–2 m depth (Rieger et al. 1972). For practical reasons we were not able to provide data for replicate watersheds. Many watershed studies share this problem of no replication (e.g. Borman & Likens 1977; Swank & Waide 1988).

Vegetation patterns in the two watersheds are closely correlated with permafrost and aspect (Figure 2). South facing slopes in LoP are dominated by deciduous forest including paper birch (*Betula papyrifera*) and quaking aspen (*Populus tremuloides*). Deciduous forest covers 43% of the LoP watershed. The remaining forest (covering 37% of LoP) is coniferous and consists of white spruce (*Picea glauca*) and black spruce (*Picea mariana*). Black spruce is found on the most shaded soils. In HiP stands of black spruce and white spruce cover 61% of the watershed area. Stands of mixed paper birch and white spruce cover only 8% of the watershed. Soils on cold slopes with more extensive permafrost are Ester silt loams (Histic lithic cryaquepts, Rieger et al. 1972). The soils of spruce and deciduous forest on warmer slopes are described as Gilmore silt loams (Typic cryothents).

The remaining 16% of LoP and 18% of HiP are ridge tops and valley bottoms and support shrublands. Permafrost can be found underneath most of the valley bottom and the water table is often visible as ponding in depressions between low hummocks. The valley bottoms are dominated by dwarf birch (*Betula glandulosa*), bog blueberry (*Vaccinium uliginosum*) and willow (*Salix* spp.) with ribbons of thin-leaf alder (*Alnus tenuifolia*) along stream edges. The soils in the valley bottoms are described as Karshner silt loams (Pergelic cryaquepts, Rieger et al. 1972). The silts comprising both valley and hill soils are very fine with 75–90% falling within the 20 μm–60 μm particle size range (White et al. 1995). There is very little clay present in these soils (<10%) and it is poorly weathered (White et al. 1995). Cation exchange capacities are also very low, ranging from 30–65 milliequivalents/100 g in organic horizons to 1–5 milliequivalents/100 g in mineral horizons (Troth et al. 1976). Troth et al. (1976) found that soils in CPCRW beneath spruce forest and deciduous forest had few differences in pH, percent C, percent N or C/N ratio. Base saturation of Ca, Mg and K was significantly higher in deciduous forest soils than in spruce forest soils however.

Snowmelt in LoP typically occurs in early April, slightly before snowmelt in HiP and marks the beginning of the soil thaw (Ray 1988). The depth of active layer (depth of soil above permafrost that freezes and thaws seasonally) in the Yukon-Tanana uplands typically reaches a maximum in August or September (Van Cleve et al. 1981).

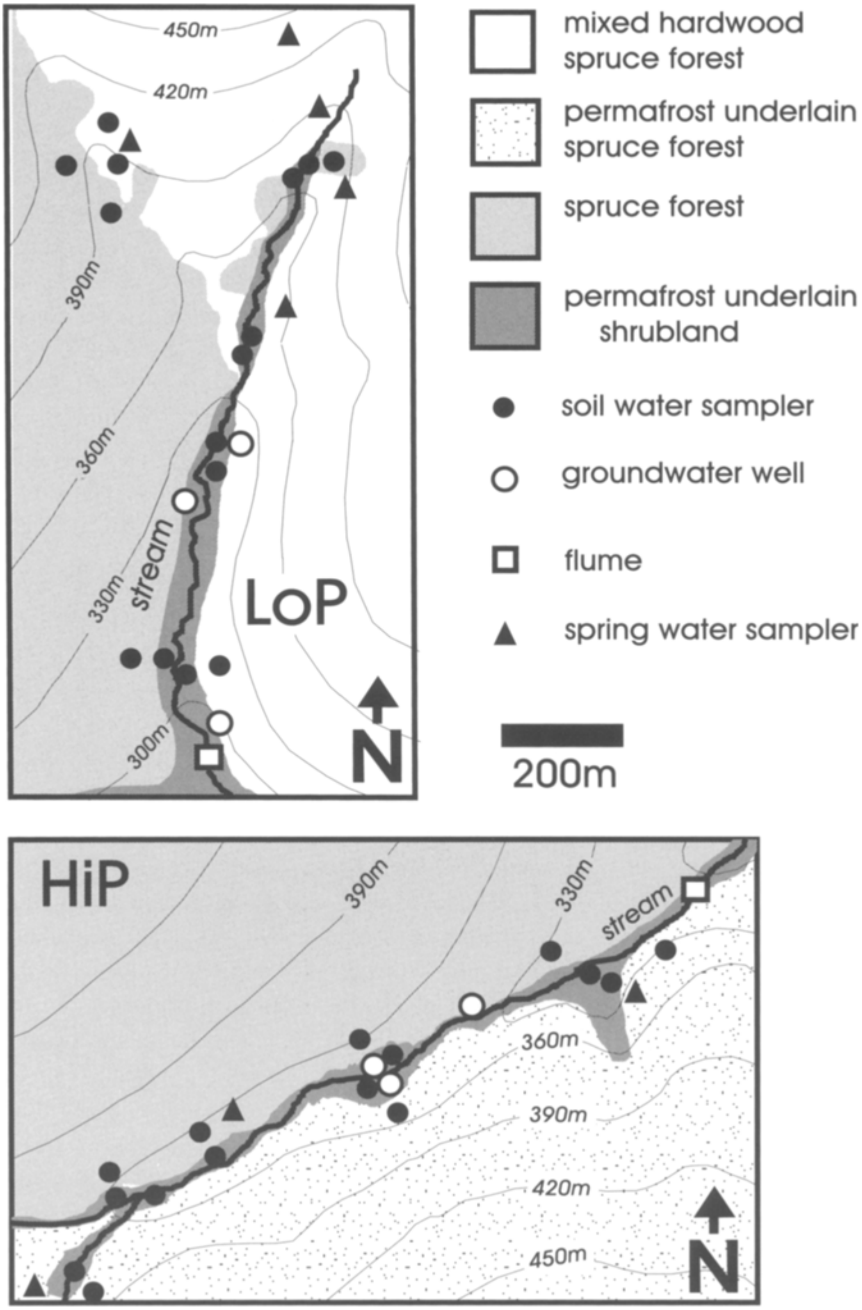

Figure 2. Topography, permafrost, vegetation types and the distribution of sampling sites in LoP (low permafrost) and HiP (high permafrost) study watersheds. Only the instrumented parts of the watersheds are shown.

Hydrological measurements and stream water sampling

We measured stream discharge using permanently installed Parshall flumes near the mouth of each watershed (Figure 1). Stream stage was recorded every 60 minutes on dataloggers (Easylogger, Omnidata Inc.) and converted to discharge using rating equations. We estimated precipitation for the two watersheds from four automatic measurement stations (gravimetric collectors, locations shown in Figure 1). We used precipitation data from 1979–1985 (Ray 1988) and 1994–1995 (CPCRW unpublished data) to calculate regressions for precipitation as a function of elevation. Hypsometric curves for LoP and HiP, which give the percent area of the watersheds at given elevation bands (Ray 1988), were used to estimate annual precipitation for each watershed. We collected stream water samples biweekly at the LoP and HiP flumes from 21 March (before snowmelt) to 15 October 1995 (after the streams froze over). We collected additional stream samples every two days during snowmelt or rainstorm recessions. Three grab samples of 500 ml were collected in each stream at each sampling in acid-washed polyethylene bottles. We had to clear both flumes of ice prior to snowmelt in April to allow sampling and accurate discharge estimates. Ice in the two streams in 1995 was not extensive enough to allow stream water to bypass the flumes. Automatic monitoring of streamflow was curtailed in October as water in the flumes froze. We visited the streams during the winter of 1994–1995 on two occasions (15 November 1994 and 5 December 1994). On these visits we cleared the ice in the flumes, estimated discharge and took stream water samples.

Soil and groundwater sample collection

We collected soil water and groundwater samples from 44 sites throughout LoP and HiP (Figure 2) between 7 July and 17 September 1995. Soil water samples were collected in PVC suction lysimeters with ceramic cups (Model 1900, Soil Moisture Inc., Santa Barbara, U.S.A.) installed to 0.3–0.5 m depth. We installed replicate suction lysimeters (n = 8) on each side of each of LoP and HiP and in valley bottoms (n = 8) in August 1994 and June 1995. Lysimeters were primed with 50–85 cbars of vacuum pressure 24 hours prior to sampling. We collected spring water in suction lysimeters installed immediately uphill from seeps. Suction lysimeters were used instead of standpipes to sample springs because the ceramic caps were useful in excluding fine silt from water samples.

When we installed the lysimeters in June 1995, we measured the active layer depth directly from the lysimeter hole. We measured permafrost depths again on 15 September 1995 by probing with a 1 m long steel rod. At some sites the presence of rocks in the soil prevented probing to the full 1 m depth however.

We drilled three well holes in each of LoP and HiP with a portable gasoline-powered drill. Wells were positioned at least 10 m from the stream bed. Well holes were lined with 51 mm ID PVC well casing with a 1.5 m long screened section at the bottom. The slit width of the well screen was 0.25 mm and samples were invariably silty. Depths from the soil surface (from the top of the A horizon) to the top of the screen on the well casing (the sample depth) ranged from 2.6 to 4.1 m in LoP and 2.1 to 4.6 m in HiP. Drilling revealed what appeared to be horizontal ice lenses separated by unfrozen soil layers 0.1 m to 0.5 m thick in the valley bottoms of both watersheds. The water sampled from the wells apparently originated from between or below these lenses. We collected soil, spring and well water samples on eight occasions between 7 July and 4 September 1995.

We collected all water samples in acid-washed 500-ml polyethylene bottles. Samples were placed in a cooler with ice packs, returned to Fairbanks, and refrigerated at 4 °C ($<$12 hours) before we filtered them using precombusted glass fiber filters (Gelman AE, 0.4 μm nominal pore size, ashed at 450 °C for 3 hours). The first 200 ml of each sample was used to rinse the filter and discarded. The remaining 300 ml was filtered and subsampled for analysis. We froze subsamples for anion and dissolved organic analysis immediately after filtering and those for dissolved inorganic nitrogen (DIN) analysis were preserved with 5 $\mathrm{mg \cdot l^{-1}}$ Hg as phenyl mercuric acid and refrigerated. Subsamples for total aluminum (Al), calcium (Ca), total magnesium (Mg), total sodium (Na), total manganese (Mn), total silicon (Si) and total potassium (K) analysis were acidified with concentrated nitric acid (Baker Ultrapure) to pH $<$2 and refrigerated.

Chemical analyses

We measured conductivity using a Hach Conductivity/TDS meter. Sulfate (SO_4^{2-}) concentrations in water samples were measured on a Dionex ion chromatograph. We found that chloride concentrations were close to the detection limits of the ion chromatograph ($<$100 μgL^{-1}) and concentrations in some deionized water blanks exceeded stream water concentrations. We measured Al, Ca, Mg, Na, Mn and Si using direct current argon plasma (DCAP) spectroscopy (Willard et al. 1981) on a Beckman Spectrascan model VI spectrometer. K was measured by atomic absorption using a Perkin-Elmer 5000 atomic absorption spectrophotometer with an air acetylene flame (American Public Health Association 1985). We analyzed NH_4^+-N and NO_3^--N using standard phenate and cadmium reduction methods respectively (American Public Health Association 1985) on a Technicon autoanalyzer. Dissolved reactive phosphorus (DRP) was measured using a stannous chloride method but concentrations were below detection limits for many of the soil and groundwater samples (<0.001 ppm). We froze subsamples for DOC and dissolved organic nitrogen DON measurements, packed them in insulated boxes and express-mailed them to the University of New Hampshire, Durham for analysis. DOC measurements were made using high temperature catalytic oxidation (HTCO) with infra-red detection of CO_2 (Shimadzu TOC 5000, Cauwet 1994). Total dissolved nitrogen (TDN) was measured by HTCO (Shimadzu TOC 5000) with chemiluminescent detection of nitrogen (Antek 720C, Merriam et al. 1996). DON was calculated as the difference between TDN and DIN.

Data analysis

We obtained rain and snow chemistry data from the National Atmospheric Deposition Program (NADP unpublished data), which has a collection site for wet precipitation at the center of CPCRW within 6 km of the LoP and HiP watersheds. Dry deposition has not been measured at this NADP site.

We estimated concentrations in stream water (volume-weighted means) and exports of solutes on a daily basis by linear interpolation between dates on which stream water was sampled. Estimates of total annual exports of dissolved materials include a linear interpolation of chemistry and discharge between the last fall sample (17 October 1995) and the first spring sample (27 March 1995). To obtain this estimate we had to treat the fall sample as if it were from the fall previous (1994). As the streams freeze over, the contribution of water from shallow soils declines gradually and the proportion of water originating from groundwater increases to close to 100% (Ray 1988). We obtained a set of samples from the stream on 15 November 1994 and 5 December 1994. These samples were analyzed for conductivity and DOC. They were within 10% of DOC concentrations and conductivity estimated by linear interpolation. To test the validity of interpolating over the winter months we performed a similar interpolation on stream chemistry data collected year-round at CPCRW (Ray 1988). Using Ray's (1988) data, we interpolated between October and April samples and then compared the estimate from the interpolation with the actual stream concentrations on the days that were sampled during this period. Chemical concentrations estimated by interpolation were all within 10% of actual sample concentrations. Because of the low flow variance in streams during the winter months (Ray 1988), the errors introduced by interpolation over the season when the stream was iced over were unlikely to be any greater than those introduced by interpolation between sampling dates of much higher frequency during the summer.

We grouped sample sites into site types. In the LoP watershed these were: spruce forest ($n = 4$), deciduous forest ($n = 3$), valley bottom ($n = 8$), springs

($n = 4$), wells ($n = 3$) and stream (one sampling point, three replicate samples). In the HiP watershed these were: spruce forest ($n = 4$), spruce forest underlain by permafrost ("frozen", $n = 4$), valley bottoms ($n = 8$), springs ($n = 3$), wells ($n = 3$) and stream (one sampling point, three replicate samples). Chemical data from different site types would not conform to assumptions of normality even following transformation. We ranked data from all sites on all dates together and determined differences between site types by running a one-way analysis of variance on the ranked data (Conover & Inman 1981) followed by a Bonferroni pairwise comparison (Neter et al. 1990).

We regressed chemical concentrations in stream water against log stream discharge. Differences between regression slopes and intercepts between the two streams or between spring and summer seasons were calculated using *t*-tests (Glantz & Slinker 1990).

Contributions of stormflow and baseflow were estimated using a graphical separation technique (Linsley et al. 1982, Figure 3). We also estimated the relative proportions of stormflow and baseflow in the streams using conductivity data. Streamflow is conceptualized as a mixture of stormflow and baseflow. We estimated relative proportions of stormflow and baseflow using a two component mixing-model (Pearce et al. 1986):

$$Q_1 + Q_2 = 1,$$
$$Q_1 = [(C_t - C_2)/(C_1 - C_2)],$$

where Q_1 and Q_2 represent the relative proportions of stormflow and baseflow. C_t represents the average conductivity of streamflow whereas C_1 and C_2 represent the conductivity of the stormflow and baseflow. Stormflow conductivity was estimated from the conductivity of precipitation measured at the NADP site. Baseflow conductivity was estimated from stream water conductivity before breakup.

Results

For those sites on the north facing slopes of HiP ($n = 4$) and those on the valley bottoms of LoP ($n = 8$) and HiP ($n = 8$) permafrost depth averaged 0.44 m on June 20 1995 ($n = 20$, variance = 0.11). There were no significant differences ($p > 0.05$) between permafrost depths in these three areas as determined by a one-way Analysis of Variance. We did not find any permafrost within 1 m of the soil surface (top of the A horizon) at any other site. When we probed the sites again on September 15 1995 we did not find permafrost within 1 m of the soil surface at any site although the presence of permafrost may have been masked by rock fragments in the soil.

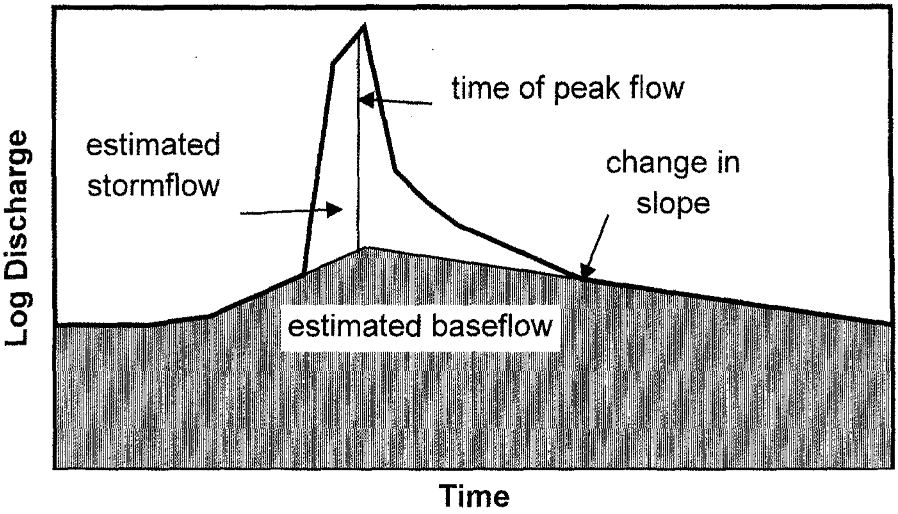

Figure 3. Graphical method of hydrograph separation. The final recession component following the hydrograph peak is extended in a line from the change in slope to the time of the peak. The line continues down to the pre-peak flow. The area below this line is considered to be baseflow, the area above is considered to be stormflow (after Linsley et al. 1982).

Streamflow was much more variable in HiP (mean = 45 $L \cdot s^{-1}$, variance = 2638) than in LoP (mean = 31 $L \cdot s^{-1}$, variance = 85) based on daily averages from 18 April through 13 September 1995 ($n = 148$). HiP had higher stormflow peaks than LoP (Figure 4). Baseflow was more sustained in LoP and made up a higher proportion of total streamflow based on graphical separation (Table 1). Mixing-model estimates of baseflow (Table 1) were within 10% of estimates from graphical separation. The mixing-model was based on the following data. Pre-snowmelt (early April) stream water conductivities were 90 $\mu S \cdot cm^{-1}$ in LoP and 100 $\mu S \cdot cm^{-1}$ in HiP and were used as estimates of baseflow conductivity for each stream. Average annual conductivity (volume-weighted) was 81 $\mu S \cdot cm^{-1}$ in LoP and 61 $\mu S \cdot cm^{-1}$ in HiP. Snow and rainwater conductivity averaged 3.3 $\mu S \cdot cm^{-1}$ and was used to approximate stormflow. Actual stormflow conductivity is likely to be somewhat higher than that of pure rainwater because of interaction of rainwater with soil and organic matter. Conductivity of ponded water measured in LoP and HiP on 20 May 1995 averaged 12 $\mu S \cdot cm^{-1}$ (variance = 2.3 $\mu S \cdot cm^{-1}$). If stormflow conductivity were 10 to 15 $\mu S \cdot cm^{-1}$ this would give baseflow estimates of only 1–2% lower than those in Table 1.

The major chemical difference between the two streams was that HiP had higher average DOC but lower dissolved mineral concentrations than LoP (Table 2). DOC concentrations in HiP were twice as high as those in LoP. Stream water from HiP had lower conductivity than stream water from

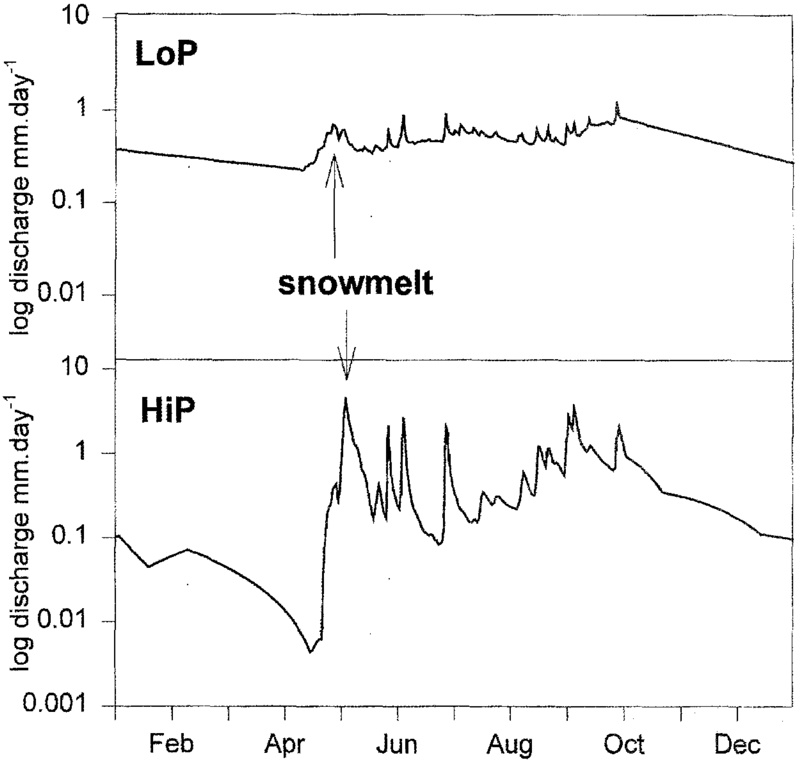

Figure 4. Hydrographs for low permafrost (LoP) and high permafrost (HiP) watersheds during the 1995 calendar year. The hydrograph peaks caused by snowmelt are indicated. Those peaks following are caused by rainfall events.

LoP suggesting that LoP stream water had greater ionic concentrations. Both streams were dominated by Ca and Mg. Ca and Mg concentrations were highly correlated with conductivity ($R^2 = 0.99$ for the sum of Ca and Mg). The main anion in these streams is HCO_3^- (Ray 1988) which was not measured in this study. LoP had twice the Mg concentration of HiP but half the Mn and Al concentrations; these differences are also seen in total exports (Table 3). The greatest absolute (rather than proportional) difference in inorganic concentrations measured between the two streams was in Ca and Mg. There was little difference in dissolved nitrogen concentrations (DON and DIN) between the two streams. An average of 82% (LoP) and 85% (HiP) of the dissolved nitrogen in the streams was in organic form.

Estimated streamflow exports were all greater (at least one order of magnitude) than estimated wet deposition, with the exception of DIN (Table 3). DIN wet deposition inputs in 1995 were lower than stream exports; however, the average annual wet deposition of DIN for the years 1993 to 1996

Table 1. Hydrologic summary for study watersheds for the 1995 calendar year. Snowpack estimates were not available. Baseflow estimates from both graphical separation ("graphical") and mixing model methods are shown for the year for low permafrost (LoP) and high permafrost (HiP) watersheds.

Parameter	Watershed	
	LoP	HiP
Total annual runoff (mm)	167	152
% baseflow (graphical)	96	66
% baseflow (mixing model)	90	60
Total precipitation (mm)	480	465
Total runoff/total precipitation	0.35	0.33
Snowmelt runoff (mm)[a]	16	27
% baseflow	81	18
Summer runoff (mm)[b]	76	92
% baseflow	95	60
Summer rainfall	407	393
Summer runoff/rainfall	0.19	0.23

[a] 12 April–18 May

[b] 19 May–30 September

for the NADP site was 0.22 kg·ha^{-1}·yr^{-1} which was approximately equal to exports. LoP and HiP were a net source of Ca, Mg, Na, K and SO_4^{2-}.

During snowmelt, DOC concentrations and dissolved minerals (represented by conductivity) were strongly correlated with log stream discharge in both streams but in summer, strong correlations between discharge and concentration were generally restricted to HiP (Figure 5(d) & (h)). DOC was positively correlated with flow in both streams. In contrast, conductivity (which is representative of dissolved minerals) was negatively correlated with discharge. Nitrogen concentrations did not correlate well with discharge (Figure 5(i–p)). There was a negative correlation between snowmelt discharge and DIN in LoP (Figure 5(m)). In this case there was a pronounced hysteresis in which concentrations in the rising stream water were higher than concentrations in receding stream water. This suggests DIN was "flushed out" of the soil by snowmelt. There was also a weak correlation between summer discharge and DON in HiP (Figure 5(l)).

The presence of strong correlations implies that stream chemistry was largely controlled by discharge rate. When we regressed chemical concentrations in stream water against log discharge (Table 4), the chemistry of HiP was generally better predicted by discharge (21 significant regressions, $P <$ 0.05) than that of LoP (12 significant regressions, $P < 0.05$). Regression

Table 2. Mean concentrations (mgL^{-1}) of selected compounds in streamwater from low permafrost (LoP) and high permafrost (HiP) watersheds. Means are volume-weighted.

Species	Watershed	
	LoP	HiP
Na	1.06	1.01
NH_4-N	0.01	0.01
K	0.61	0.57
Ca	10.1	8.9
Mg	2.70	1.40
Al	0.06	0.11
Mn	0.028	0.057
SO_4^{2-}	1.37	1.35
NO_3-N	0.14	0.12
DRP*	0.001	0.001
Si	3.90	3.66
DOC	2.31	6.45
DIN	0.15	0.13
DON	0.68	0.71
DOC:DON	3.38	9.04
Conductivity $\mu S \cdot cm^{-1}$	81	61

* Dissolved reactive phosphorus (below detection limits)

equations were typically steeper in LoP than in HiP (Table 4). Chemical behavior in LoP was more predictable during snowmelt (10 significant regressions, $P < 0.05$) than in summer (2 significant regressions, $P < 0.05$). In contrast, there were 9 significant regressions ($P < 0.05$) in HiP during snowmelt and 13 during summer.

Na, Ca, Mg, SO_4^{2-} and conductivity were all negatively correlated with discharge rate whereas DOC:DON ratios, DOC and Al were positively correlated. Al concentrations were positively correlated with DOC concentrations in HiP for both snowmelt ($R^2 = 0.74$, $P < 0.01$) and summer ($R^2 = 0.69$, $P < 0.01$) and in LoP during snowmelt ($R^2 = 0.77$, $P < 0.01$) but not summer. Most compounds measured maintained either a positive or negative relationship with discharge from snowmelt into the summer. The exceptions, in both streams, were Si and NO_3^--N which were diluted by snowmelt (negative slope) but increased in concentration with increasing runoff later in the summer (positive slope). K, DRP and Mn concentrations were poorly predicted by discharge.

Table 3. Annual (1995) stream output estimates compared with depositional inputs. All values are in kg·ha^{-1}·yr^{-1} unless otherwise stated. LoP and HiP indicate the low and high permafrost watersheds respectively. Deposition data were obtained from the National Atmospheric Deposition Program site in CPCRW. Dashes (–) indicate data that were not collected.

	LoP output	HiP output	Wet deposition input
Ca	15.1	13.6	0.06
Mg	4.0	2.1	0.016
Na	1.6	1.5	0.127
K	0.9	0.9	0.029
NH_4^+-N	0.02	0.01	0.070
SO_4^{2-}	2.1	2.1	0.48
NO_3^--N	0.21	0.19	0.09
DRP	0.002	0.002	–
Si	5.83	5.59	–
Mn	0.042	0.086	–
Al	0.097	0.174	–
DIN	0.23	0.20	0.16
DON	1.02	1.08	–
DOC	3.45	9.77	–
DOC:DON	3.38	9.04	–

Site type had a significant effect ($P < 0.001$) on the concentrations of all chemicals measured with the exception of DRP as determined by ANOVA on ranked data. Data for DRP was below detection limits for soil and groundwater sites. Pairwise comparisons showed that soil water collected from 0.3 to 0.5 m beneath deciduous forest and spruce forest had lower ($P < 0.05$) Na, Mg, Ca, SO_4^{2-} and conductance than well water (Table 5, Figure 6(a)). The concentrations of total Na, Mg, Ca and SO_4^{2-} over sampling sites were strongly correlated with conductivity ($R^2 = 0.99$, $P < 0.001$ for the sum of Ca, Mg and Na, $R^2 = 0.83$, $P < 0.01$ for SO_4^{2-}). Al concentrations were positively correlated with DOC concentrations over all sites ($R^2 = 0.20$, $P < 0.001$). Significant differences in K and Mn concentrations did not follow an explainable pattern.

We expected stream water, spring water and valley bottom water to be mixtures of soil water and well water and therefore to have intermediate conductivity. While this was true, water sampled from the LoP valley bottom had a conductivity as high as that of well water from LoP (Figure 6(a)). Valley bottom water in LoP and HiP had Ca and Mg concentrations (Table 5) that

Table 4. Regression coefficients for stream chemical concentrations versus log stream discharge for 1995. The regression equation is [y] = M log(Q) = C where [y] is the concentration ($\text{mg}\cdot\text{l}^{-1}$), M is the slope, log(Q) is $\log_{10}$ of discharge ($\text{l}\cdot\text{ha}^{-1}\cdot\text{s}^{-1}$) and C is the intercept. LoP and HiP indicate the low and high permafrost watersheds respectively. Standard errors (SE) are in parentheses. The significance of the regressions (P) are denoted * ($P < 0.05$), ** ($P < 0.01$) and *** ($P < 0.001$). Significant ($P < 0.05$) differences between the regression equations for LoP and HiP within either the snowmelt or summer seasons are indicated by parentheses around the stars for significance (**). Differences between the snowmelt or summer season within LoP or HiP are indicated in the "Season" column. For each stream, $n = 10$ for snowmelt and $n = 32$ for summer seasons.

		Snowmelt (12 April–18 May)				Summer(19 May–30 September)				Season
Species	Site	Slope (±SE)	Intercept (±SE)	R^2	*P*	Slope (±SE)	Intercept (±SE)	R^2	*P*	*P*
Na	LoP	−0.73 (±0.18)	−0.13 (±0.23)	0.68	(**)	ns	–	–	–	–
	HiP	−0.24 (±0.02)	0.55 (±0.04)	0.93	(***)	−0.35 (±0.05)	0.75 (±0.06)	0.59	***	***
NH_4^+-N	LoP	ns	–	–	–	ns	–	–	–	–
	HiP	ns	–	–	–	−0.03 (±0.01)	−0.02 (±0.01)	0.26	**	–
K	LoP	ns	–	–	–	ns	–	–	–	–
	HiP	ns	–	–	–	−0.22 (±0.05)	0.27 (±0.06)	0.37	***	–
Ca	LoP	−5.97 (±0.92)	1.7 (±1.21)	0.82	***	−4.05 (±1.86)	5.60 (±2.25)	0.13	*	*
	HiP	−4.00 (±0.38)	3.88 (±0.71)	0.92	***	−5.98 (±0.28)	3.27 (±0.33)	0.93	***	***
Mg	LoP	−2.11 (±0.59)	−0.18 (±0.78)	0.58	(**)	−1.43 (±0.48)	1.24 (±0.59)	0.22	**	–
	HiP	−0.44 (±0.05)	0.77 (±0.10)	0.89	(***)	−0.54 (±0.06)	0.94 (±0.07)	0.70	***	***
Al	LoP	0.25 (±0.04)	0.39 (±0.06)	0.77	(***)	ns	–	–	–	–
	HiP	0.06 (±0.02)	0.25 (±0.04)	0.53	(***)	0.08 (±0.01)	0.16 (±0.01)	0.56	***	***
Mn	LoP	0.35 (±0.15)	0.58 (±0.19)	0.39	*	ns	–	–	–	–
	HiP	ns	–	–	–	ns	–	–	–	–

Table 4. Continued.

		Snowmelt (12 April–18 May)				Summer(19 May–30 September)				Season
Species	Site	Slope (±SE)	Intercept (±SE)	R^2	P	Slope (±SE)	Intercept (±SE)	R^2	P	P
SO_4^{2-}	LoP	–	–	–	–	ns	–	–	–	–
	HiP	−0.49 (±0.04)	−0.16 (±0.07)	94	***	−0.29 (±0.06)	0.2 (±0.07)	40	***	**
DRP	LoP	ns	–	–	–	ns	–	–	–	–
	HiP	ns	–	–	–	ns	–	–	–	–
NO_3-N	LoP	−0.16 (±0.03)	−0.1(±0.03)	0.82	***	ns	–	–	–	–
	HiP	ns	–	–	–	0.03 (±0.01)	0.16 (±0.01)	18	*	**
Si	LoP	−1.99 (±0.36)	0.93 (±0.48)	0.77	(***)	3.1 (±1.06)	7.76 (±1.29)	0.21	(**)	***
	HiP	−0.53 (±0.13)	2.19 (±0.24)	0.65	(**)	0.61 (±0.13)	4.66 (±0.15)	0.43	(***)	***
DOC	LoP	14.9 (±2.97)	24.1(±4.02)	0.74	(***)	ns	–	–	–	–
	HiP	4.22 (±0.81)	14.5 (±1.6)	0.75	(***)	4.19 (±0.32)	9.33 (±0.37)	0.85	***	***
DIN	LoP	−0.23 (±0.04)	−0.17 (±0.05)	0.81	***	ns	–	–	–	–
	HiP	ns	–	–	–	ns	–	–	–	–
DON	LoP	ns	–	–	–	ns	–	–	–	–
	HiP	ns	–	–	–	0.16 (±0.04)	0.9 (±0.04)	0.38	***	–
DOC:DON	LoP	ns	–	–	–	ns	–	–	–	–
	HiP	4.66 (±1.80)	16.67 (±3.34)	0.43	*	3.78 (±0.38)	9.55 (±0.44)	0.76	***	***
Conduc.	LoP	−1.41 (±0.22)	−1.16 (±0.29)	0.82	(***)	ns	–	–	–	–
μS·cm^{-1}	HiP	−0.26 (±0.04)	0.2 (±0.07)	0.85	(***)	−0.29 (±0.02)	0.36 (±0.03)	0.82	***	***

Table 5. Average concentrations of selected chemicals in soil water, groundwater and stream water. Samples were from low permafrost (LoP) and high permafrost (HiP) watersheds for the period 7 July to 17 September 1995. Standard errors (SE) are in parentheses. "Spruce", "deciduous" and "frozen" refer to soil water collected from black spruce forest, deciduous forest, and black spruce forest with permafrost.

		Average concentration ($mg{\cdot}l^{-1}$)							
	Site	Al	Ca	K	Mg	Na	SO_4^{2-}	Si	Mn
LOP	Spruce	0.30 (0.17)	3.6 (3.4)	0.64 (0.21)	0.90 (0.59)	0.84 (0.37)	0.21 (0.20)	5.27 (1.24)	0.25 (0.07)
	Deciduous	0.53 (0.35)	3.8 (2.4)	1.24 (0.57)	1.15 (0.70)	1.33 (0.76)	0.14 (0.20)	6.72 (0.65)	0.61 (0.39)
	Valley	0.11 (0.21)	14.2 (8.1)	0.73 (0.16)	3.78 (1.90)	1.36 (0.67)	0.60 (0.47)	5.02 (0.55)	1.17 (0.29)
	Springs	0.03 (0.02)	9.3 (3.8)	0.76 (0.27)	2.53 (0.48)	1.20 (0.29)	0.78 (0.33)	4.62 (0.64)	0.03 (0.01)
	Wells	0.45 (0.81)	13.4 (3.8)	1.29 (0.24)	3.59 (0.76)	1.86 (0.56)	1.64 (0.29)	4.78 (1.06)	0.52 (0.11)
	Stream	0.03 (0.02)	10.5 (0.6)	0.60 (0.11)	2.95 (0.22)	0.94 (0.15)	1.35 (0.29)	4.01 (0.36)	0.02 (0.01)
HIP	Spruce	0.37 (0.16)	3.6 (6.8)	0.40 (0.15)	0.47 (0.10)	0.81 (0.34)	0.18 (0.46)	4.77 (1.43)	0.15 (0.03)
	Frozen	0.45 (0.19)	3.6 (4.7)	0.32 (0.32)	0.69 (0.51)	0.73 (0.30)	0.33 (0.59)	3.87 (0.59)	0.12 (0.02)
	Valley	0.08 (0.05)	9.9 (3.8)	0.56 (0.12)	1.73 (0.33)	1.40 (0.30)	1.75 (0.92)	4.06 (0.41)	0.09 (0.02)
	Springs	0.07 (0.05)	22.3 (12.4)	0.89 (0.13)	2.55 (1.27)	1.63 (0.32)	4.08 (3.57)	4.62 (0.66)	0.38 (0.09)
	Wells	0.41 (0.44)	13.4 (4.0)	1.36 (0.48)	2.21 (0.36)	1.79 (0.43)	1.65 (0.43)	4.47 (0.77)	0.05 (0.39)
	Stream	0.09 (0.07)	10.2 (2.4)	0.54 (0.09)	1.52 (0.37)	1.13 (0.17)	1.56 (0.52)	3.97 (0.34)	0.02 (0.01)

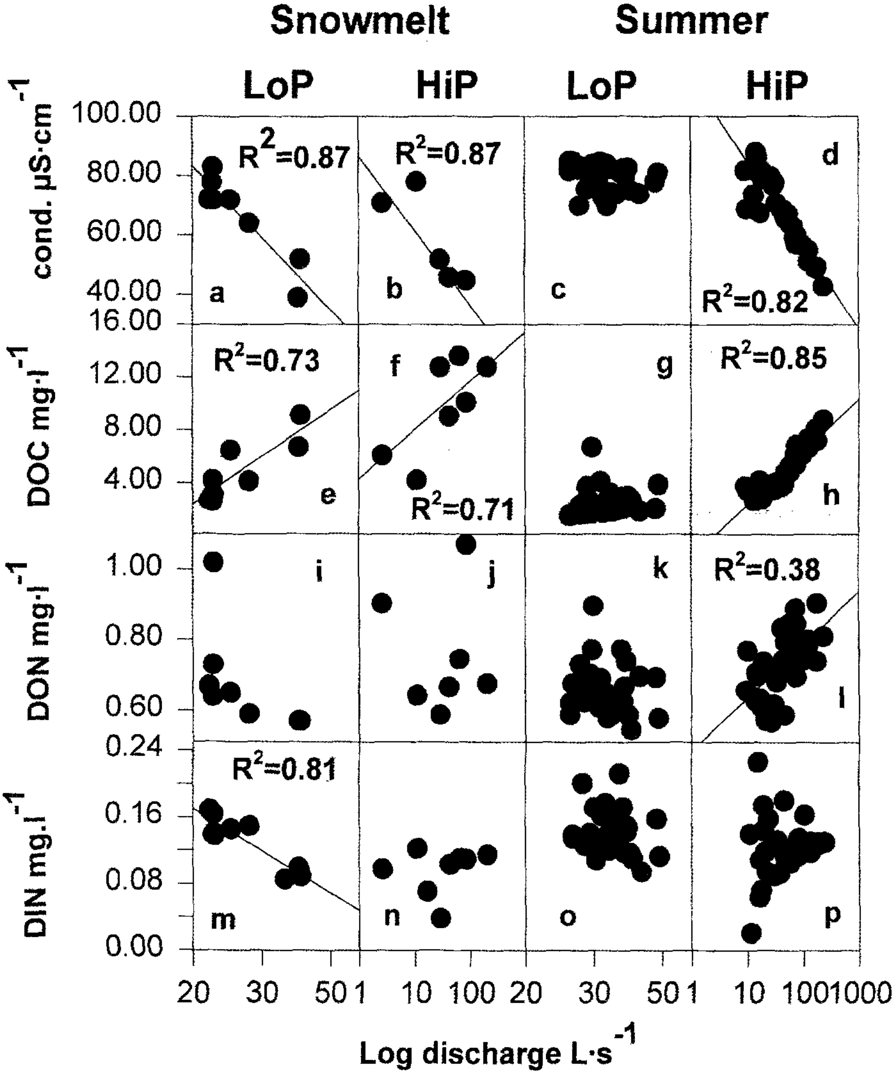

Figure 5. Scattergraphs of stream chemistry versus log stream discharge. R^2 values and lines of best fit are shown for significant ($P < 0.05$) regressions. Note differing range of discharge values for high permafrost (HiP) and low permafrost (LoP) watersheds.

were statistically similar ($P > 0.05$) to well water implying a similar origin. Spring water in HiP had a conductance, Ca:Mg ratios and SO_4^{2-} concentrations significantly higher ($P < 0.05$) than well water suggesting that spring water and well water originate from different sources.

DOC concentrations were significantly higher ($P < 0.05$) in soil water samples from spruce and deciduous forests than in samples from other types

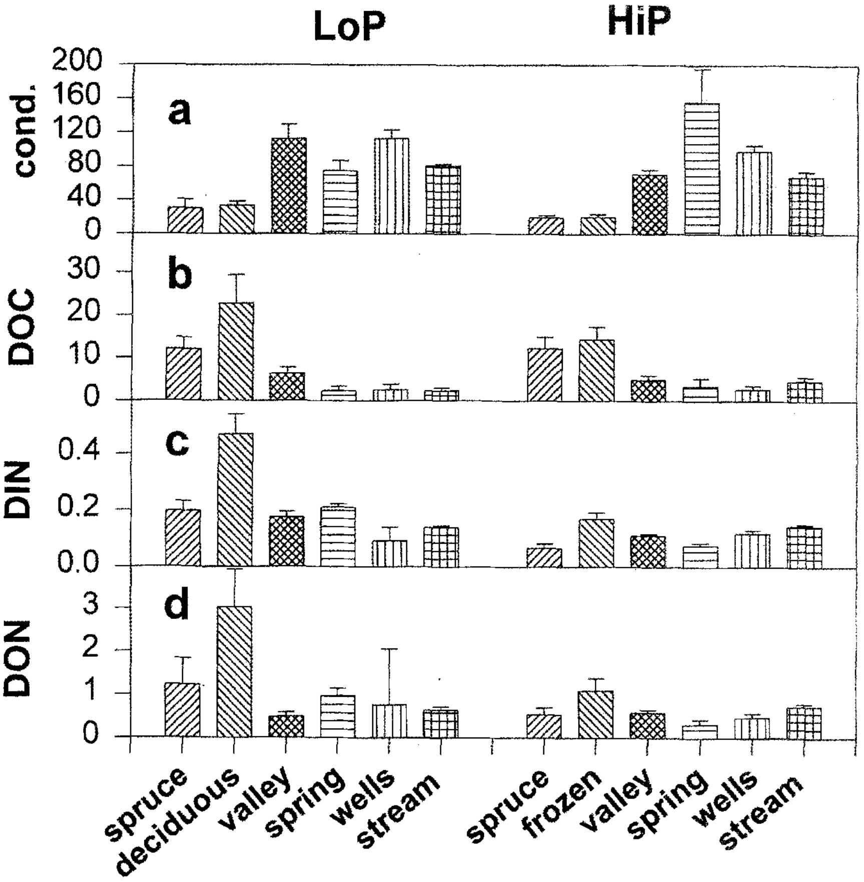

Figure 6. Mean conductivity (Cond. $\mu S \cdot cm^{-1}$), DOC ($mg \cdot l^{-1}$), DON ($mg \cdot l^{-1}$) and DIN ($mg \cdot l^{-1}$) by site type. "Spruce", "deciduous" and "frozen" refer to soil water collected from black spruce forest, deciduous forest, and black spruce forest with permafrost in low permafrost (LoP) and high permafrost (HiP) watersheds.

of sites (Figure 6b). There were no significant differences among the remaining types of sites however. Deciduous forest in LoP had significantly higher ($P < 0.05$) DIN and DON concentrations than soil water from any other type of sample site in this study (Figure 6(c) and (d)).

DOC:DON ratios in the streams were lower ($P < 0.05$) than they were in water from the forest floor (Figure 7(a)). Deciduous forest had much lower DOC:DON ratios than spruce forest with and without permafrost. NO_3^--N:NH_4^+-N ratios were significantly lower in samples from valley bottoms in LoP and in spruce in HiP than they were in samples from either stream (Figure 7(b)). There were no significant differences between other sites and the streams however.

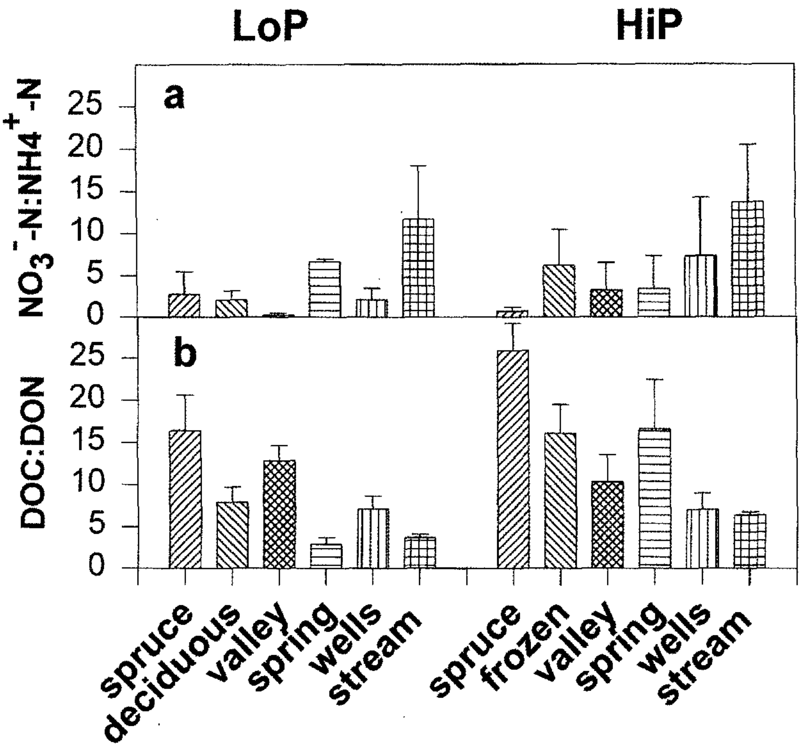

Figure 7. DOC:DON and NO_3^--N:NH_4^+-N ratios in water samples by sampling site. "Spruce", "deciduous" and "frozen" refer to soil water collected from black spruce forest, deciduous forest, and black spruce forest with permafrost in low permafrost (LoP) and high permafrost (HiP) watersheds.

There were no consistent temporal trends in the chemistry of water samples from wells and soil suction lysimeters. Individual sites did show temporal trends but when these were averaged within site types (wells, springs, valley bottoms, forest types) they could not be resolved.

Discussion

The primary differences in stream chemistry between LoP and HiP can be attributed to the presence of permafrost. The main chemical differences between stream water in LoP and HiP was that HiP had higher DOC and lower dissolved mineral concentrations than LoP. Soil water and groundwater sampled within the rooting zone was high in DOC but low in dissolved minerals (Figure 6), and it is apparent that DOC concentration dropped and dissolved minerals increased as this water moved deeper into the mineral horizons and underlying schist bedrock. Stormflow made up a significantly larger proportion of total streamflow in HiP than LoP (Table 1). This stormflow had consistently high DOC concentrations and low dissolved mineral concentrations as determined by the chemistry of stream water at high flow rates (Figure 5). Together, these factors support the hypothesis that the more extensive permafrost in HiP confines runoff to the upper organic horizons where it becomes enriched with DOC and generates stormflow with high DOC concentrations. The permafrost inhibits the movement of water through deeper mineral horizons where it can accumulate dissolved minerals.

During snowmelt, DOC, Ca and Mg concentrations and conductivity in LoP and HiP streams exhibited similar dynamics, but this similarity faded in summer. This was probably due to the lack of significant variation in discharge in the summer in LoP (Figure 4). Upper soil horizons in LoP were frozen during snowmelt and this appeared to have a similar effect on the dynamics and chemistry of spring runoff in LoP as did permanent freezing (permafrost) on the dynamics and chemistry of overall runoff in HiP. As the active layer thawed in LoP, variance in discharge and chemical concentrations narrowed. This is further support for the importance of permafrost in regulating DOC and major ion concentrations in the stream.

The reduction of DOC concentration as water moved down through mineral soils in LoP and HiP is consistent with the behavior of DOC in other studies. DOC immobilization by sorption, observed in mineral soils (McDowell & Wood 1984; Qualls & Haines 1991) and in stream bed gravel (McDowell 1985; Fiebig & Lock 1991) in temperate watersheds has been invoked to explain the depletion of DOC in groundwater and baseflows in boreal watersheds (Wallis et al. 1981; Koprivnjak & Moore 1992). The texture of the soils in our study area have a very low clay content (White et al. 1995) which is not normally considered a favorable condition for effective sorption (Jardine et al. 1989). Apparently the fineness of the silt in these soils or some other property still enables the retention of DOC.

Water retained in upper soil horizons appeared to be relatively dilute especially in spruce forest soils (Figure 6, Table 5). The dilution of major inorganic ions (Ca, Mg, Na and SO_4^{2-}) in the streams at high flows (Figure 5, Table 4) is most likely caused by runoff of this water from upper soil. The higher CEC in upper organic horizons of the soils (Troth et al. 1976) may also be responsible for diluting ions in the runoff contributing to storm flows.

As water moves below the upper soil horizons becoming groundwater, Ca, Mg, Na and SO_4^{2-}accumulate as determined from spring and well water (Table 5). The same is likely to be true of HCO_3^-, the major anion in these streams. Although it was not measured in this study, the concentration dynamics of HCO_3^- typically follows those of Ca, Mg, Na and SO_4^{2-} in

CPCRW streams (Ray 1988). This deeply percolating water is the source of baseflows in LoP and HiP as is reflected by the increase in conductivity and dissolved minerals at low flows (Figure 5).

K, Mn, Al and Si concentrations in stream water are poorly predicted by the dynamics of streamflow, in contrast to Ca, Mg, Na and SO_4^{2-}, and appear to be influenced by biological uptake or chemical processes. It is harder to generalize about the effects of permafrost on K, Mn, Al and Si fluxes. Total Al concentrations appear to follow those of DOC, suggesting that most dissolved Al occurs in chelated form with DOC (Evans 1986; Pohlman & McColl 1988). Si concentrations in stream water are diluted by snowmelt pulses in spring but these concentrations increase during summer rainstorms. Active layers are deeper in the summer allowing greater water movement through mineral soil and fractured bedrock. High flows through this matrix and possible changes in soil pH from increased decomposition may result in greater dissolution of Si during the summer. In summer, K concentrations may be influenced by plant uptake. Van Cleve et al. (1983) estimated the annual K requirement for taiga forest to differ between 0.13 $g{\cdot}m^{-2}$ in black spruce and 2.89 $g{\cdot}m^{-2}$ in deciduous forests. There is no information on the nutritional requirements of the vegetation in valley bottoms (dominantly dwarf birch and blueberry). We estimated K exports from both LoP and HiP to be 9 $g{\cdot}m^{-2}$. Ca and Mg requirements are of a similar magnitude to those for K (Van Cleve et al. 1983) but stream exports are one to two orders of magnitude higher (Table 3). Plant uptake therefore is likely to have much less influence on stream concentrations of Ca and Mg than K.

Nitrogen cycling dynamics in watershed soils appear to be more important than hydrology in regulating N flux in LoP and HiP. The similarity in DIN and DON concentrations between the two streams and the absence of any relationship between dissolved nitrogen and discharge differs from the behavior of DOC, SO_4^{2-}, Ca, Mg and Na. The high concentrations of DIN and DON beneath deciduous forest compared to other sample sites could be a reflection of higher primary productivity and larger pools of soluble nutrients in soils of deciduous forest compared to those of other vegetation types (Van Cleve et al. 1981). Higher DIN and DON concentrations may also be due to the higher densities of N fixing alder (*Alnus crispa*) beneath deciduous forest as compared to spruce forest in CPCRW (Troth et al. 1976).

Despite higher DIN concentrations under deciduous forest in LoP, DIN concentrations in the two streams were not significantly different suggesting that soils in LoP may be more efficient in intercepting DIN than HiP. The consistently low NO_3^--N:NH_4^+-N ratios in the valley bottom of LoP (Figure 7b) may indicate greater denitrification in this watershed compared to HiP. The fact that discharge does not fluctuate as widely or as quickly in LoP as in

HiP (Figure 4) implies that the valley bottoms of LoP do not drain as quickly as those in HiP. Slow-moving water beneath a relatively unchanging water table is more likely to create anaerobic conditions conducive to denitrification. The greater interception of DIN in LoP compared to HiP could also be driven by differences in plant uptake or microbial immobilization rates.

The dominance of DON over DIN in stream water is characteristic of old growth temperate forest watersheds with low inputs of anthropogenic nitrogen (Hedin et al. 1995) and appears to be a characteristic of CPCRW. Approximately 70% of total N exports from HiP and LoP was in the form of DON. Atmospheric inputs of nitrogen are relatively low at CPCRW (Table 3) which has two orders of magnitude less nitrate and sulfate in wet deposition than temperate forests (Likens & Borman 1995). This is primarily because of the great distance of Interior Alaska from major sources of industrial pollution. Average DIN concentrations in LoP and HiP are typical for small unpolluted streams in North America (0.2 $\mathrm{mg \cdot l^{-1}}$; Allan 1995). Input-output estimates (Table 3) suggest that LoP and HiP may be net nitrogen sinks which would be typical of forests in mid-successional phase or those receiving low inputs of nitrogen in precipitation (Aber et al. 1989; Hedin et al. 1995).

Nitrate and ionic concentrations in both streams are higher than those reported for first order Arctic watersheds which are completely underlain by permafrost (Lock et al. 1989; Everett et al. 1996). The difference in nitrate is possibly because of lower primary productivity in the arctic tundra resulting in smaller pools of DIN. Lower ionic concentrations in arctic streams are presumably because contact between runoff and mineral soil horizons is more limited by permafrost in the arctic. DOC concentrations in these first order streams are higher on average than LoP and HiP. Oswood et al. (1996) report average annual DOC concentrations in Imnavait Creek of 11 and 13 $\mathrm{mg \cdot l^{-1}}$. Lock et al (1989) report average DOC concentrations of between 5–9 $\mathrm{mg \cdot l^{-1}}$ in three first order streams from the same region. These Alaskan data suggest that DOC concentrations may be a function of permafrost distribution.

Water draining through spruce forest soils underlain by permafrost appeared to have higher DOC:DON ratios than that draining through deciduous forest soils without permafrost. Total C:N ratios and total C in soil do not vary significantly between the spruce and deciduous forest types in CPCRW (Troth et al. 1976). A likely reason for the high DOC:DON ratios in water draining spruce forest is that primary productivity and rates of nutrient cycling are usually lower in these cooler, permafrost soils resulting in smaller pools of dissolved nitrogen (Van Cleve et al. 1981). If DOC and DON adsorbed equally we would expect that DOC:DON ratios in stream water would be similar to those in uplands. Instead DOC:DON ratios are significantly lower in stream water than in upland soils (Figure 6(e)). This

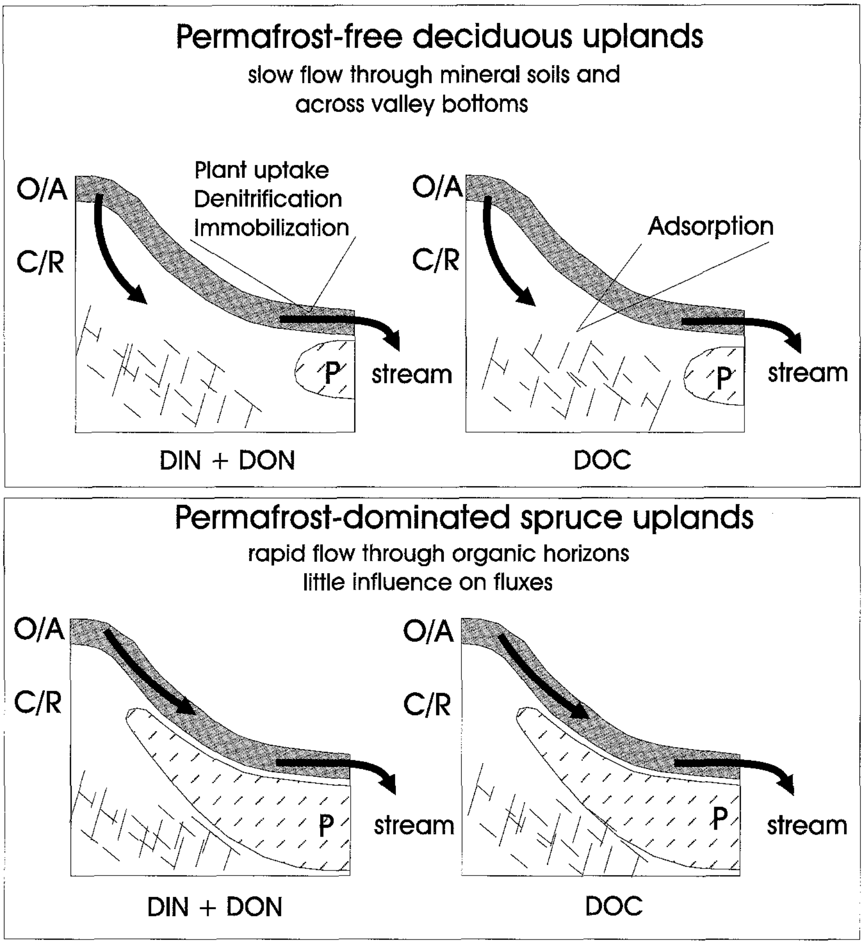

Figure 8. Hypothesized model for the effect of permafrost on N (DON and DIN) and DOC. We hypothesize that permafrost reduces the passage of water through mineral soils and speeds its passage through soil/stream ecotones in valley bottoms. This results in less effective modulation of DIN, DON and DOC concentrations. "O/A" indicates organic and A horizons; "C/R" indicates mineral horizons and bedrock and "P" indicates permafrost. Arrows indicate pathways of water movement and are not to scale.

implies either that sorption or mineralization of DOC is more efficient than that for DON or that additional DON is being introduced into the runoff on its way to the stream or in the stream itself. Qualls and Haines (1991) found that decomposition played a relatively minor role in the reduction of DOC and DON concentrations in comparison to adsorption in a temperate forest hydrologic profile, suggesting that changes in the DOC:DON ratios seen in LoP and HiP are unlikely to be due to mineralization.

Stream chemistry in permafrost-dominated watersheds appears to be more tightly coupled with the chemistry of organic horizons in the upper soil than is the chemistry of streams draining permafrost-free watersheds. Permafrost-underlain soils generate runoff rapidly and confine it to upper soil horizons (Slaughter & Kane 1979; Woo 1986; Shaver et al. 1990). Consequently, there is less opportunity for the chemistry of this runoff to be modulated by transit through mineral soil. In contrast the chemistry of runoff from permafrost-free soils is controlled by contact between water and mineral soils. In addition, the slower, more consistent runoff from permafrost-free soils is more likely to sustain waterlogged areas in valley bottoms and provide more opportunities for plant uptake of nutrients and denitrification.

From a watershed management perspective, presence of permafrost appears to compromise the buffering properties of mineral soil and stream-soil ecotones. Streams in permafrost-dominated watersheds are likely to be more sensitive to nutrient inputs to upland and valley bottom soils than those in permafrost-free watersheds. Nutrient inputs from natural sources such as fire or human sources such as septic waste systems seem more likely to be flushed through the watershed and into stream water in permafrost-dominated watersheds than they are in permafrost-free watersheds. Changes in the distribution of permafrost due to natural or anthropogenic climate change therefore have the potential to change both the biogeochemistry of high latitude watersheds and the abilities of the taiga landscape to modulate the movement of dissolved material from upland soils into streams. In particular we predict that loss of permafrost would result in decreased hydrologic losses of DOC and DON and increased losses of Ca and Mg.

Acknowledgements

Major funding for this study was provided by the Bonanza Creek Long-Term Ecological Research (LTER) program of the National Science Foundation (DEB-9211769) and by a Dissertation Improvement Grant (DEB-9520594), also from the National Science Foundation. MacLean was also supported by a Global Change Research Fellowship and a Graduate Resource Fellowship from the University of Alaska Fairbanks. We thank Rich Boone, Joy Curley, Allen Doyle, Lola Oliver, Tim Quintal, and Josh Schimel for help and advice regarding laboratory analyses and, especially, Jeff Merriam for organic carbon and organic nitrogen analyses. Field assistance was provided by Ron Hartzman, Dave Kaplan, Scott Smidt, Darrin Lee, George Rogers, and Zach Mondry. A rig for drilling wells was loaned by Mike Lily and instruction provided by Scott Ray. Larry Hinzman provided key help in experimental design and hydrological modeling. We would also like to thank two anonymous reviewers for their constructive criticism and review.

References

Aber JD, Mellilo JM, Nadelhoffer KJ, Pastor J & Boone RD (1989) Factors controlling N cycling and N saturation in northern temperate forest ecosystems. Ecol. Appl. 1: 303–315

Allan JD (1995) Stream Ecology. Chapman and Hall, London, U.K.

American Public Health Association, American Water Works Association, and Water Pollution Control Federation (1985) Standard Methods for the Examination of Water and Wastewater, 16th edn. Washington DC, U.S.A.

Cauwet G (1994) HTCO method for dissolved organic carbon analysis in seawater: Influence of catalyst on blank estimation. Marine Chemistry 47: 55–64

Van Cleve K, Barney R & Schlentner R 1981 Evidence of temperature control of production and nutrient cycling in two interior Alaska black spruce ecosystems. Can. J. For. Res. 11: 258–273

Van Cleve K, Oliver L, Schlentner R, Viereck LA & Dyrness CT (1983) Productivity and nutrient cycling in taiga forest ecosystems. Can. J. For. Res. 13: 747–766

Conover WJ & Inman RL (1981) Rank transformations as a bridge between parametric and nonparametric statistics. Am. Statistician 35: 124–129

Buttle JM & Fraser KE (1992) Hydrochemical fluxes in a high arctic wetland during spring snowmelt. Arct. Alp. Res. 24: 153–164

Evans A Jr (1986) Effects of dissolved organic carbon and sulfate on aluminium mobilization in forest soil columns. Soil Sci. Soc. Am. J. 50: 1576–1578

Everett KR, Kane DL & Hinzman LD (1996) Surface water chemistry and hydrology of a small arctic drainage basin. In: Reynolds JF & Tenhunen JD (Eds) Landscape Function and Disturbance in Arctic Tundra (pp 185–201). Ecological Studies 120, Springer, Berlin, Germany.

Fiebig DM & Lock MA (1991) Immobilization of dissolved organic matter from groundwater discharging through the stream bed. Freshwater Biol. 26: 45–55

Ford TR (1973) Precipitation runoff characteristics of the Caribou Creek research watershed near Fairbanks, Alaska. MS Thesis, University of Alaska Fairbanks, AK, U.S.A.

Ford TE & Naiman RJ (1989) Groundwater – surface water relationships in boreal forest watersheds: Dissolved organic carbon and inorganic nutrient dynamics. Can. J. Fish. Aquat. Sci. 46: 41–49

Glantz SA & Slinker BK (1990) Primer of Applied Regression and Analysis of Variance. McGraw-Hill, New York, U.S.A.

Haugen RK, Slaughter CW, Howe KE & Dingman SL (1982). Hydrology and climatology of the Caribou-Poker creeks research watershed, Alaska. CRREL Report No 82-26, US Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, NH, U.S.A.

Hedin LO, Arnesto JJ & Johnson AH (1995) Patterns of nutrient loss from unppolluted, old growth temperate forests: evaluation of biogeochemical theory. Ecology 76: 493–509

Jardine PM, Weber NL & McCarthy JF (1989) Mechanisms of dissolved organic carbon absorption on soil. Soil Sci. Soc. Am. J. 53: 1378–1385

Kling GW, O'Brien WJ, Miller MC & Hershey AE (1992) The biogeochemistry and zoogeographyof lakes and rivers in arctic Alaska. Hydrobiol. 240: 1–14

Koprivnjak JF & Moore TR (1992) Sources, sinks and fluxes of dissolved organic carbon in subarctic fen catchments. Arct. Alp. Res. 24: 204–210

Koutz FR & Slaughter CW (1973) Equivalent latitude and a permafrost environment. CRREL Technical Note, US Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, NH, U.S.A.

Likens GE & Borman FH (1995) Biogeochemistry of a Forested Ecosystem, 2nd edn. Springer Verlag Publishers, New York, U.S.A.

Linsley RK, Kohler MA & Paulhus JCH (1982) Hydrology for Engineers. McGraw Hill Publishers, New York, U.S.A.

Lock MA & Ford TE (1986) Colloidal and dissolved organic carbon dynamics in undisturbed boreal forest catchments: A seasonal study of apparent molecular weight spectra. Freshwater Biol. 16: 187–195

Lock MA, Ford TE, Fiebig DM, Miller MC, Hullar M, Kaufman M, Vestal JR, Peterson BJ & Hobbie JE (1989) A biogeochemical survey of rivers and streams in the mountains and foot-hills province of arctic Alaska. Arch. Hydrobiol. 115: 499–521

McDowell WH & Wood T (1984) Podzolization: Soil processes control dissolved organic carbon concentrations in stream water. Soil Science 137: 23–31

McDowell WH (1985) Kinetics and mechanisms of dissolved organic carbon retention in a headwater stream. Biogeochemistry 1: 329–352

McNamara JP, Kane DL & Hinzman LD (1997) Hydrograph separations in an arctic watershed using mixing-model and graphical techniques. Water Resour. Res. 33: 1707–1720

Merriam J, McDowell WH & Currie WS (1996) A high temperature catalytic oxidation technique for determining total dissolved nitrogen. Soil. Sci. Soc. Am. J. 60: 1050–1055

Naiman, RJ (1982) Characteristics of Sediment and Organic Carbon Export from Pristine Boreal Forest Watersheds. Can. J. Fish. Aquat. Sci. 39: 1699–1718

National Atmospheric Deposition Program (unpublished data) Internet address for data (URL): http://nadp.nrel.colostate.edu/NADP

National Weather Service (unpublished data) Climate summary for Fairbanks. Internet address for data (URL): http://www.wrcc.sage.dri.edu

Neter J, Wasserman W & Kutner MH (1990) Applied Linear Statistical models, 3rd edn. Irwin, Boston, U.S.A.

Oswood MW, Irons JG III & Schell DM (1996) Dynamics of dissolved and particulate carbon in a tundra stream in arctic Alaska. In: Reynolds JF & Tenhunen JD (Eds) Landscape Function and Disturbance in Arctic Tundra (pp 275–287). Ecological Studies 120, Springer, Berlin, Germany

Pearce AJ, Stewart MR & Sklash MG (1986) Storm runoff generation in humid headwater catchments 1: Where does the water come from? Water Resour. Res. 22: 1263–1271

Peterson BJ, Hobbie JE & Corliss TL (1986) Carbon flow in a tundra stream ecosystem. Can. J. Fish. Aquat. Sci. 43: 1259–1270

Peterson BJ, Corliss T, Kriet K & Hobbie JE (1992) Nitrogen and phosphorus concentrations and export for the upper Kuparak River on the North Slope of Alaska in 1980. Hydrobiol. 240: 61–69

Pohlman AA & McColl JG (1988) Soluble organics from forest litter and their role in metal dissolution. Soil Sci. Soc. Am. J. 52: 265–271

Qualls RG & Haines BL (1991) Geochemistry of dissolved organic nutrients in water percolating through a forest ecosystem. Soil. Sci. Soc. Am. J. 55: 1112–1123

Ray SR (1988) Physical and chemical characteristics of headwater streams at Caribou-Poker creeks research watershed. MS thesis, University of Alaska Fairbanks, Fairbanks, Alaska

Rieger S, Furbush, CE, Schoephorster DB, Summerfield H & Geiger LC Jr (1972) Soils of the Caribou-Poker creeks research watershed, Interior Alaska. CRREL Technical Report 236, US Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Hanover, NH, U.S.A.

Slaughter CW & Kane DL (1979) Hydrologic role of shallow organic soils in cold climates In: Canadian Hydrology Symposium 79 – Cold Climate Hydrology. National Research Council of Canada (pp 380–389)

Shaver GR, Nadelhoffer KJ & Giblin AE (1991) Biogeochemical diversity and element transport in a heterogeneous landscape, the North Slope of Alaska. In: Turner MG & Gardner RH (Eds) Quantitative Methods in Landscape Ecology; The Analysis and Interpretation of Landscape Heterogeneity (pp 105–125). Ecological Studies 82, Springer, Berlin, Germany

Swank WT & Waide JB (1988) Characterization of baseline precipitation and nutrient budgets for control watersheds. In: Swank WT & Crossley DA (Eds) Forest Hydrology and Ecology at Coweeta (pp 57–79). Springer Verlag, New York, U.S.A.

Troth JL, Deneke FJ & Brown LM (1976) Upland aspen/birch and black spruce stands and their litter and soils properties in interior Alaska. For. Sci. 22: 33–44

Viereck LA, Dyrness CT, Van Cleve K & Foote MJ (1983) Vegetation, soils and forest productivity in selected forest types in interior Alaska. Can. J. For. Res. 13: 703–720

Wallis PM, Hynes H & Telang SA (1981) The importance of groundwater in the transportation of allochthonous dissolved organic matter to streams draining a small mountain basin Hydrobiologia 79: 77–90

White JD, Koepke BE & Swanson DK (1995) Soil survey of North Star Area, Alaska, Interim Report. US Department of Agriculture Natural Resources Conservation Service

Willard HH, Merrit LL, Dean JA & Settle FA Jr (1981) Instrumental Methods of Analysis, 6th edn. D Van Nostrand Publishers, New York, U.S.A.

Woo MK (1986) Permafrost hydrology in North America. Atmos. Ocean 24: 201–234

Woo MK & Winter TC (1993) The role of permafrost and seasonal frost in the hydrology of northern wetlands in North America. J. Hydrol. 141: 5–31