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Changing hydrologic connectivity due to permafrost thaw in the lower Liard River valley, NWT, Canada

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Abstract:

Flows from river basins in northwestern Canada have been rising in the last two decades as a result of climate warming. In the wetland-dominated basins that characterise the southern margin of permafrost, permafrost thaw and disappearance, and resulting land-cover change, is occurring at an unprecedented rate. The impact of this thaw on runoff generation in headwater basins is poorly understood. Permafrost thaw has the potential to fundamentally alter the cycling and storage of moisture inputs in this region by altering the type and relative proportions of the major land-cover types, such as peat plateaus, channel fens and flat bogs. This paper examines streamflow changes in the four Water Survey of Canada gauged river basins (152–2050 km^2) in the lower Liard River valley, Northwest Territories, Canada, a region where permafrost thaw has produced widespread loss of forest and concomitant expansion of permafrost-free wetlands. Annual runoff in the lower Liard Valley increased by between 112 and 160 mm over the period of 1996–2012. The Mann-Kendall non-parametric statistical test and the Kendall-Theil robust line were used to ascertain changes in streamflow. Historical aerial photographs from 1977 and high-resolution satellite imagery (WorldView 2) from 2010 were used to measure the rate and pattern of permafrost thaw in a representative 6 km^2 area of Scotty Creek. Permafrost thaw-induced land-cover change is both increasing the adjacency between runoff producing and transmitting land cover types and transforming certain land covers that store water into ones that produce runoff. This land-cover change was found to be the single most important factor (37–61 mm) contributing to the observed increase in river discharge. Other contributing factors include increases in plateau runoff contributing areas (20–32 mm), increases in annual effective precipitation depth (18–30 mm), contribution of water from the melt of ice within permafrost (9 mm) and increases in baseflow (0.9–6.8 mm). Although runoff has significantly ($p < 0.05$) increased in all four basins, the largest increases are in basins with a relatively high cover of flat bogs.

KEY WORDS permafrost thaw; hydrology; wetlands; runoff

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INTRODUCTION

The hydrological regime of Arctic and subarctic rivers is changing in response to anthropogenic global warming. Increases in discharge have been noted in northern Canada (Déry *et al*., 2009; St. Jacques and Sauchyn, 2009) and in several large rivers in Eurasia (Peterson *et al*., 2002; McClelland *et al*., 2004; Smith *et al*., 2007). Déry and Wood (2005) reported that rivers flowing into Hudson Bay and the Labrador Sea experienced declines in discharge between 1964 and 2003; however, Déry *et al*. (2009) show a reversal of this trend when expanding the observational period by just 4 years (1964–2007). This exemplifies the high inter-annual variability of streamflow records and also illustrates the influence of high streamflows in recent years on long-term trends. Reasons for increasing streamflow in Arctic and subarctic rivers vary. It has been suggested that changes to runoff patterns are closely related to global climate warming patterns (Peterson *et al*., 2002). Changes to atmospheric oscillation cycles and moisture distribution stemming from climate change have been attributed as possible explanations for observed increases in discharge (Déry *et al*., 2009).

General circulation models coupled with hydrological models have predicted further increases of river discharge to the Arctic Ocean in the coming decades (Manabe *et al*., 2004; Milly *et al*., 2005; Wu *et al*., 2005). These models predict an increase in atmospheric moisture in northern latitudes in the coming decades, which will lead to increases in river discharge (Manabe *et al*., 2004). Climate warming has also intensified permafrost thaw, which could have a variety of implications for streamflow in permafrost covered basins. One example is additional moisture inputs

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from the conversion of ground ice to water (McClelland *et al.*, 2004). St. Jacques and Sauchyn (2009) speculate that the thawing of permafrost may lead to the reactivation of groundwater systems, thereby leading to an increase in stream flow. Altering the spatial distribution of permafrost may change existing flow paths of groundwater in permafrost-dominated regions. For example, Smith *et al.* (2005) show that the amount of ponded water (i.e. lakes) nearly doubles in permafrost regions as opposed to adjacent non-permafrost regions because of restricted subsurface flow pathways. Karlsson *et al.* (2012) suggest that thawing permafrost will alter the connection pathways between surface and subsurface water. Thawing permafrost could change the hydrological regime of a region from one of predominantly surface flow to one where subsurface flow would dominate, thereby impacting the basin hydrograph (Walvoord and Striegl, 2007; St. Jacques and Sauchyn, 2009; Rowland *et al.*, 2010).

St. Jacques and Sauchyn (2009) found that nine of 23 rivers analysed in the Northwest Territories (NWT), Canada, show significant ($p \leq 0.1$) increases in stream flow over a 30-year period. Of these nine rivers, five are clustered in the southwest NWT on or near the southern edge of discontinuous permafrost (lower Liard River valley). St. Jacques and Sauchyn (2009) suggest that these increasing discharge rates may be the result of reactivated groundwater systems resulting from permafrost thaw. These authors, using winter flows as a proxy to estimate baseflows, show significant ($p \leq 0.1$) increases to winter flows in 20 of the 23 basins. Although the basins in the lower Liard River valley have shown significant increases in winter flows, these flows account for less than 7% of total annual flows, indicating that groundwater reactivation is not the primary driver of rising stream flows in this region. We suggest that it is not through the activation of groundwater pathways that permafrost thaw is increasing stream flows but rather through increased surface and near-surface hydrological connection of wetlands resulting from permafrost-thaw-induced land-cover change.

The climate in northwestern Canada has been warming at a rate faster than anywhere else on Earth (Johannessen *et al.*, 2004). This has led to widespread permafrost thaw and a northward migration of the southern limit of discontinuous permafrost in the Mackenzie River valley (Kwong and Gan, 1994). The rate of permafrost thaw is greatest in the zone of discontinuous permafrost, because in this zone, permafrost bodies warm and thaw from both vertical energy fluxes from the ground surface and horizontal fluxes from adjacent permafrost-free terrains (McClymont *et al.*, 2013). Permafrost in this southern margin is also warm ($>-0.5\,°C$) and relatively thin ($<10\,m$) (Burgess and Smith, 2000), compared with the continuous permafrost zone. The loss of permafrost often leads to the complete conversion of forest ecosystems to wetlands (Quinton *et al.*, 2011). Vitt *et al.* (1994) show that vegetation on undisturbed peat plateaus exerts a control on the supra-permafrost water table. The combination of canopy transpiration and an unsaturated organic layer keeps the water table at a sufficient depth to allow for the establishment of rooting zones. In areas experiencing permafrost degradation, lateral thawing of permafrost results in ground subsidence, bringing the water table closer to the ground surface (Jorgenson and Osterkamp, 2005).

Permafrost significantly influences the partitioning of precipitation into infiltration and runoff and influences drainage patterns because the surface of the permafrost bodies rise above adjacent permafrost-free terrains, thereby impounding or redirecting water. In the lower Liard River valley, permafrost exists predominantly below treed peat plateaus, which are surrounded by adjacent wetlands (i.e. flat bogs and channel fens). Quinton *et al.* (2003) suggested that each of these three land-cover types has a unique function in the basin hydrological cycle: Peat plateaus act as runoff generators owing to their small storage capacity (Wright *et al.*, 2009) and higher elevation relative to the adjacent wetlands, flat bogs are primarily storage features and channel fens convey water to the basin outlet. Given the contrasting hydrological functions of bogs, fens and plateaus, permafrost-thaw-induced changes to the landscape therefore have the potential to influence the hydrograph response of basins (Quinton *et al.*, 2003).

Image analysis of time-separated aerial photos and/or satellite images is commonly used to estimate the rate and pattern of permafrost loss using forest loss as a proxy method of estimating permafrost loss (Tutubalina and Rees, 2001; Beilman and Robinson, 2003; Chasmer *et al.*, 2010). Quinton *et al.* (2011) analysed images from a $1\,km^2$ area of interest (AOI) spanning from 1947 to 2008 at Scotty Creek, NWT, situated in the lower Liard River valley. Over this period, they found that the area underlain by permafrost decreased from 70% to 43%. Permafrost-free areas were shown to be expanding and coalescing as permafrost thaw became more prevalent. The majority (72%) of the permafrost-free area lost since 1947 transformed from wooded peat plateaus to flat bogs, indicating the importance of understanding the influence of bogs on the basin hydrograph as permafrost thaw proceeds.

The present study will attempt to establish a causal link between increasing stream flows in the lower Liard River valley and the permafrost-thaw-driven land-cover changes in that region. The objectives of this study are to (1) characterise the pattern and rate of permafrost-thaw-induced change over the period of 1977–2010, (2) analyse precipitation and basin hydrograph patterns to quantify the control that permafrost thaw has on the observed increase in stream flow and (3) examine implications of the observed permafrost thaw on the basin water balance.

STUDY SITE

This study will analyse stream flow records from four gauged, wetland-dominated basins typical of the lower Liard River valley, NWT: Birch River, Blackstone River, Jean Marie River and Scotty Creek (Figure 1). Permafrost extent in this region is sporadic discontinuous (~40% coverage) and is ice rich as a result of the high porosity of the organic soils that characterise the basins (Quinton *et al.*, 2011). The region is dominated by peatlands, overlying relatively flat and poorly drained glaciolacustrine sediments (Aylesworth and Kettles, 2000). The basins are of varying sizes (152–2050 km^2); however, they share many of the same land-cover features (i.e. channel fens, flat bogs, peat plateaus and lakes). The Fort Simpson region is characterised by long, cold winters and short, dry summers. Mean annual runoff from the four basins in the lower Liard River valley is 188 mm for the longest common period of

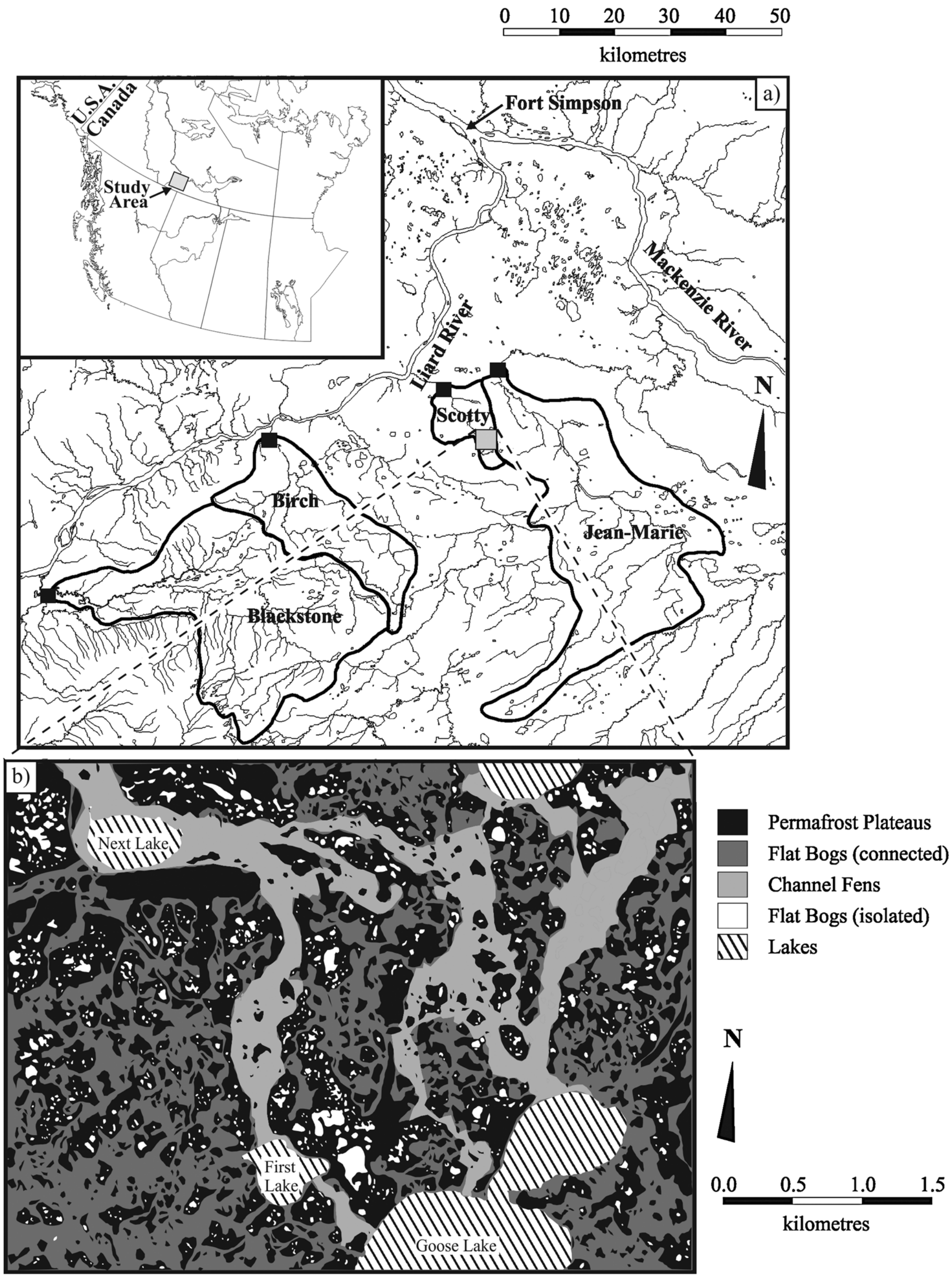

Figure 1. (a) Location of the four study basins in the Lower Liard River Valley: Blackstone River (2050 km^2), Birch River (542 km^2), Scotty Creek (152 km^2) and Jean Marie River (1310 km^2). (b) Inset presents a classified, 22 km^2 image of the Scotty Creek basin outlining the major land-cover types (Quinton *et al.*, 2008)

record (1996–2012), whereas annual precipitation at Fort Simpson is 540 mm over that same period (47% snow, 53% rain) (MSC, 2013). Hydrometric data from the four basins show an apparent increase in runoff patterns over the last 15 years (Figure 2), whereas total annual precipitation has remained relatively stable over this time.

Field work and image analysis were undertaken in the Scotty Creek (61°18′N, 121°18′W) basin. The basin features a series of wetlands that store and convey water shed from upland peat plateaus. There are three major peatland formations at Scotty Creek: permafrost peat plateaus, channel fens and flat bogs. Permafrost exists solely underneath peat plateaus and is maintained as a result of the large thermal offset provided by the insulating peat. The high porosity of near-surface peat plays an important role in the creation and preservation of permafrost (Vitt *et al*., 1994). Peat plateaus act as runoff generators, as their storage capacity is relatively low (Wright *et al*., 2009). Precipitation falling on a peat plateau will typically flow into an adjacent wetland via subsurface flow.

Runoff is slowly conveyed to the basin outlet through a series of minerotrophic channel fens, which are about 50–100 m in width. Channel fens are the most prominent feature in the basin drainage network. Hayashi *et al*. (2004) have shown through isotopic analysis that fens may become hydrologically disconnected from the drainage network during low-flow periods.

Bogs appear to be either hydrologically isolated from or seasonally connected to the channel fen system. For bogs that border channel fens, water flows directly into the fen by diffuse, mainly subsurface flow throughout the non-frozen season. Isolated bogs are bound on all sides by permafrost, which prevents flow of surface and shallow subsurface water to other wetlands. Runoff into isolated bogs from plateaus does not become available to the basin drainage network except seasonally, in response to large precipitation and melt events, and is dependent on the depth of both the frost and water tables.

Insights gained from field observations and image analysis have formed the basis of a conceptual framework (Figure 3) that describes the cycling of water within and flux from basins dominated by bogs, fens and permafrost plateaus. Direct inputs of rainfall and snowmelt reach all three peatland types, with inputs to the channel fen being conveyed directly to the basin outlet. Water entering a primary runoff contributing area of peat plateaus (Figure 4) will flow directly into a channel fen; otherwise, it will flow into either a connected or isolated bog. A connected bog, or series of connected bogs, may convey water to the channel fen (termed secondary runoff) and onto the basin outlet, whereas water entering isolated bogs will remain in storage. Some isolated bogs have ephemeral channels that cut through peat plateaus and, during periods of high moisture supply (i.e. during snowmelt or in response to large summer rain events), conduct surface and subsurface flow to downstream bogs and the channel fen. These channels are ~5–10 m wide with a ground subsidence of ~1 m, indicating a potential degradation of permafrost. The channels are characterised by dead or dying trees, plausibly due to a water table situated above or close to the ground surface. It is presumed that runoff from peat plateaus into such bogs becomes available to the basin drainage network when their ephemeral channels are active (i.e. during periods of high moisture supply); however, the volume and timing of runoff and the hydrological behaviour of the connections have not yet been studied.

Evaporatranspiration (ET) draws water from all three peatland types and is the dominant loss of water from isolated bogs. Runoff is the dominant flux from the plateaus however. For example, during the snow-free periods in spring of 2004 (29 March to 4 June) and 2005 (19 April to 8 June), runoff from a study plateau at Scotty Creek was 4.3 and 3.2 times greater than ET, respectively (Wright *et al*., 2008). The runoff ratios (runoff/precipitation) for each period were 0.87 and 0.79, respectively. ET was presumably higher during the 2005 season because of higher air

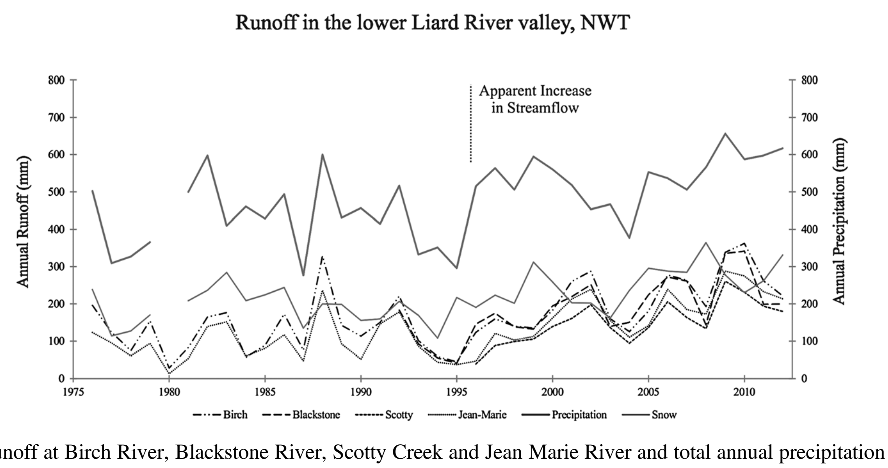

Figure 2. Total annual runoff at Birch River, Blackstone River, Scotty Creek and Jean Marie River and total annual precipitation (i.e. rain and snow) at Fort Simpson for the period 1975–2012. Both values were computed for the water year 01 October to 30 September

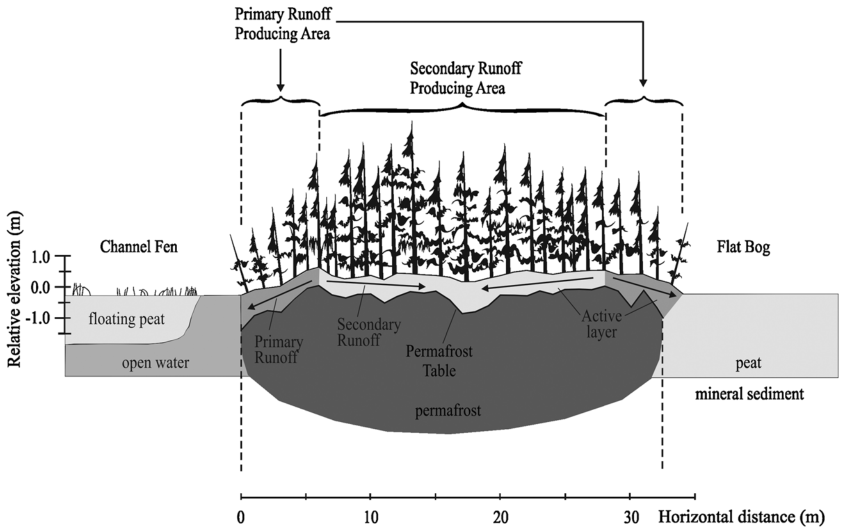

Figure 3. Cross-sectional area of a peat plateau underlain by permafrost with a channel fen and flat bog on either side. Primary runoff is shown to occur on the flanks of the peat plateau, flowing into adjacent wetlands. Secondary runoff flows into internal bogs on the peat plateau. Note the depression in the middle of the plateau: this is conducive to bog formation as it will become inundated with water and therefore become more thermally conductive resulting in a larger depression in the permafrost. This positive feedback cycle is the basis for bog creation on a peat plateau

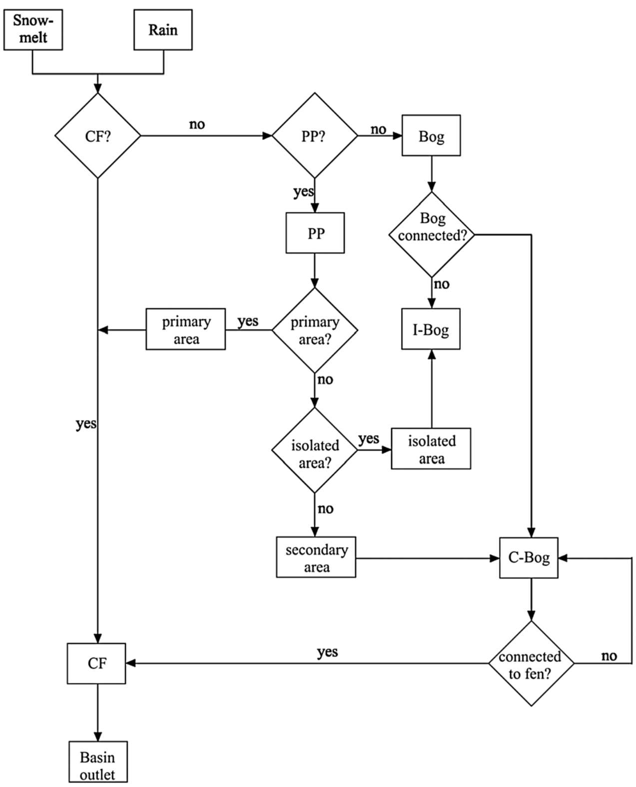

Figure 4. Conceptual model of pathways for precipitation in a wetland-dominated basin in the zone of discontinuous permafrost. Rectangles represent processes; diamonds represent questions. Primary runoff flows from a peat plateau (PP) directly into the channel fen (CF), whereas secondary runoff is routed through a series of connected bogs (C-Bog) before reaching the channel fen. Water retained as storage flows from a peat plateau into an isolated bog (I-Bog)

temperatures (and therefore a more rapid melt) during the snowmelt season (Wright *et al.*, 2008). Wright *et al.* (2008) also found that ET from an adjacent bog was about double that of the plateau, indicating higher ET rates from saturated surfaces, because of high surface moisture availability rates. The cumulative effect of ET on the basin water balance requires further study. On plateaus, the presence of permafrost prevents the interaction of supra-permafrost groundwater (within the active layer) and sub-permafrost groundwater, and although sub-permafrost groundwater systems are connected to the bogs and fens, the fluxes are minimal (Hayashi *et al.*, 2004).

Precipitation received on the sloped flanks (i.e. primary contributing area) of plateaus flows directly into the adjacent channel fen and is termed primary runoff. Unlike secondary runoff to fens, which occurs when the depression storage capacity of hydrologically connected bogs is exceeded, primary runoff to fens is direct, because the plateau flanks and corresponding hydraulic gradient [average gradient of 0.049; *n*: 15; *SD*: 0.025 (Chasmer *et al.*, in review)] slope towards the fens. Primary runoff occurs mainly as subsurface supra-permafrost flow along the margins of a peat plateau as overland flow in the primary runoff areas is rare (Wright *et al.*, 2009). Hillslope runoff from plateaus varies spatially and temporally and is strongly controlled by the depth and thickness of the saturated layer of peat, because the hydraulic conductivity of the peat decreases abruptly with depth (Quinton *et al.*, 2008). During active layer thawing, the saturated layer is bounded by the relatively impermeable and gradually lowering frost table at the lower end and by the fluctuating water table at the upper end (Wright *et al.*, 2009).

METHODOLOGY

Permafrost thaw

Because, in the type of system studied here, permafrost only exists underneath treed peat plateaus, detection of permafrost using remotely sensed imagery is relatively simple. Aerial photographs and remotely sensed imagery clearly depict the presence and boundaries of treed plateaus. The images were manually digitised in ArcGIS to delineate the boundaries of permafrost and non-permafrost terrain. Cumulative maximum errors (i.e. the worst case scenario) for this method of permafrost detection were calculated to be 8–10% over the total area of the site for the time series used in this study (Chasmer *et al*., 2010). These errors may result from orthorectification, pixel resolution, delineation errors and shadowing (Chasmer *et al*., 2010). Quinton *et al*. (2011) examined the relative change in land-cover type over a 1 km^2 AOI between 1947 and 2010. This AOI has been expanded to 6 km^2 to increase the spatially sampled portion of the basin. Two images were used for analysis: an aerial photograph from 1977 (resolution: 0.53 m) and remotely sensed imagery (WorldView 2) (resolution: 0.18 m) coupled with light detection and ranging imagery (up to ten returns per square metre) from 2010. The 1977 aerial photo was chosen because it has the highest resolution of all early (i.e. before 2000) aerial images of the basin. Total permafrost (i.e. peat plateaus) and permafrost-free areas (i.e. wetlands and lakes) were calculated for each year, as well as total wetland area and edge length that is hydrologically connected to the basin drainage network, and the area of isolated bogs. The total area of isolated bogs was calculated by finding the sum of the area of all the bogs that do not have an active connection to the basin drainage network, as determined from image analysis. It should be noted that there is potential for ephemeral channels that were not recognised in the image analysis (resulting from the presence of trees) to transmit water between these bogs during periods of high moisture supply. This analysis allows for permafrost thaw rates to be calculated over a large (6 km^2) area, as well as the growth of the basin drainage network (i.e. the area that is able to contribute flow to the basin outlet), its associated edge length and the disappearance rate of isolated bogs. As high-resolution remotely sensed imagery was only available for Scotty Creek, results will be extrapolated to the other gauged basins in the lower Liard River valley to attempt to explain increasing stream flows. Landsat imagery for all four basins indicates that these basins all exhibit similar land-cover types and distributions (Quinton *et al*., 2003), thereby making this extrapolation reasonable in this region.

The conversion of ice within the permafrost to water may contribute to increased stream flow by providing additional moisture inputs that were previously unavailable. This input is likely to be a secondary factor but should be quantified. For simplicity, permafrost depth and the porosity of peat in the basin were assumed to be homogenous and were estimated at 10 m (McClymont *et al*., 2013) and 0.8 (Quinton *et al*., 2008) respectively. It should be noted that greater permafrost thicknesses in some areas may lead to an underestimation here of the total ice volume. Assuming that the 6 km^2 AOI is representative of the entire (152 km^2) Scotty Creek basin in terms of percentage permafrost cover, we computed the total permafrost area in the basin and multiplied it by an average permafrost thickness (10 m) and peat porosity (0.8) to estimate the ice volume within the basin. This volume was multiplied by 0.91 to account for the volumetric difference between ice and water and was then normalised by dividing the volumetric input by the basin area to provide the depth of additional water. Because the annual depth of additional moisture is dependent on the initial ice content, the permafrost thaw calculations were made for each year between 1977 and 2010, on the basis of the average rate of thaw as calculated over the 6 km^2 AOI.

Analysis of precipitation and hydrograph patterns

Basin precipitation was determined from monthly and annual totals collected by Environment Canada at the Fort Simpson meteorological station (refer to Figure 1 for location). Because of the proximity (<150 km) of all four basins to this station, it is believed that this is an adequate representation of precipitation in each basin. The precipitation values have been adjusted as per Mekis and Vincent (2011). The Water Survey of Canada has been collecting discharge data (m^3/s) from the four basins dating back to 1975 (Birch), 1976 (Jean Marie), 1992 (Blackstone) and 1995 (Scotty Creek). For the purposes of this study, analysis was completed using precipitation and runoff data for the water year (1 October to 30 September) instead of the calendar year, as snow accumulation beginning in October is not recognised as runoff until thaw commences the following spring. Annual basin runoff was computed from

$$R = (Q/A) \times 1000$$

where R is annual basin runoff (mm), Q is annual discharge (m^3) and A is the basin drainage area (m^2). It should be noted that due to the low slope of all the basins (<0.0063 m/m; Table I), the boundaries of the basins are difficult to delineate. As a result, there may be some uncertainty in the basin area and, by extension, basin runoff. To gain an understanding of the partitioning of precipitation over time, the runoff ratios for each basin were calculated for the period after the increase in stream flow. Runoff ratios were calculated by dividing annual runoff by precipitation to determine the percentage of precipitation that was discharged from the basin outlet each year.

Table I. Area and per cent cover of each land-cover type at the four basins in the lower Liard River valley

Basin	Area (km^2)	Fen (%)	Bog (%)	Drainage density (km/km^2)	Slope (m/m)
Birch	542	30.7	6.5	0.373	0.0063
Jean Marie	1310	27.4	7.6	0.237	0.0034
Blackstone	1910	33.5	3.4	0.378	0.0055
Scotty	152	19.6	10.2	0.161	0.0032

Adapted from Quinton *et al.* (2003).

The non-parametric Mann-Kendall test was used to analyse long-term trends in stream flow. This test has been used in other hydrological studies analysing change in annual stream flow (i.e. Déry and Wood, 2005; Walvoord and Striegl, 2007; St. Jacques and Sauchyn, 2009). The Kendall-Theil robust line (Helsel and Hirsch, 2002) was used to create the linear equation of the regression. The Kendall-Theil robust line allows for the calculation of the magnitude of change over the period of study. Correlation was tested at the $p \leq 0.01$, $p \leq 0.05$ and $p \leq 0.1$ significance levels.

A two-step process was used to determine the year that total annual runoff began to increase. First, annual stream flow records from the Birch and Jean Marie Rivers (the two basins with long-term records) were averaged over 11-year moving windows (Déry *et al.*, 2009) to smooth out inter-annual variations. Figure 5 shows that stream flow records were relatively unchanged until the mid-1990s. Second, a regression analysis was completed to determine the year in which the increase in stream flows occurred. Multiple lines of best fit were drawn through total annual runoff, each beginning at the first year of record and ending in each year of the 1990s (i.e. 10 lines total). The point ending in the year with the minimum slope and R^2 values was chosen as the change point, as the values of each subsequent year began to influence the correlation of the data. As such, hydrograph analysis was sub-divided to include the years before the increase (1976–1995) and after the increase (1996–2012). Only the Birch and Jean Marie Rivers have datasets dating back to 1976, thereby making these the only two basins analysed during the period before the apparent increase in stream flow.

Changes in precipitation were also analysed using the Mann-Kendall test and the Kendall-Theil robust line to determine if any trends exist. Precipitation was analysed both before and after the increase in stream flow. In addition, precipitation was plotted against runoff for each basin for the periods before and after the increase in stream flow to determine how much of the increase in stream flow can be attributed to changes in precipitation. R^2 and p-values were computed for each plot in this analysis.

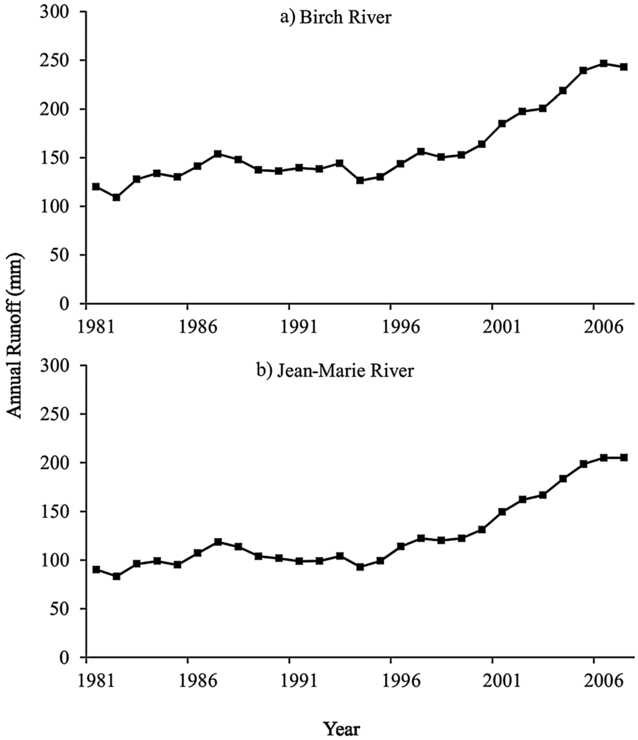

Figure 5. The 11-year moving windows of annual runoff at (a) Birch River and (b) Jean Marie River. The year indicated is the median year (i.e. 1990 represents the period from 1985 to 1995)

To be consistent with other authors, we used winter flows as a proxy to measure additional baseflow inputs in the region. Winter flows are commonly used as an approximation for baseflows (Walvoord and Striegl, 2007; St. Jacques and Sauchyn, 2009). However, these methods cannot ascertain the exact source of the winter flows, so caution must be exercised when making this assumption. Winter baseflow was calculated as all flows occurring from 01 January to 31 March (Walvoord and Striegl, 2007; St. Jacques and Sauchyn, 2009). These values were converted to millimetre using the same method as in the Section on Analysis of Precipitation and Hydrograph Patterns. This yielded the cumulative depth (mm) of winter baseflow expected during the 3-month period from 01 January to 31 March. As baseflow occurs year-round (i.e. not exclusively during this winter period), these values were multiplied by 4 to give annual baseflow contributions (mm). These results were again analysed using the Mann-Kendall test and Kendall-Theil robust line to determine annual and total baseflow additions for each basin over the observed period.

Implications of permafrost thaw on basin water balance

As outlined in the work by Quinton *et al.* (2003), each land-cover feature in the Scotty Creek basin has a distinct hydrological function. It has been shown that permafrost thaw causes the relative proportion of these land-cover features to change (Quinton *et al*., 2011). As each land-cover type exhibits a different hydrological function (in terms of the movement and storage of water), it can be surmised that permafrost thaw would result in a change in the runoff ratio of a basin. Therefore, we wanted to obtain a range of plausible runoff ratios for each basin prior to the increase in streamflow.

The Birch and Jean Marie River basins are the only two basins in the lower Liard River valley with long-term hydrometric data records (1976–2012). To find a range of acceptable runoff ratios from these basins prior to the increase in stream flow, all runoff ratios were calculated for the period of 1976–1995. These values were then sorted, and the top and bottom 25% of values were trimmed. This resulted in a range of plausible runoff ratios that would be expected prior to the increase in stream flows. Runoff ratios vary between basins and are influenced by the per cent cover of different land-cover features (Quinton *et al*., 2003). Therefore, runoff ratios calculated at the Birch and Jean Marie River basins cannot be assumed to be the same for Scotty Creek and Blackstone River. To calculate runoff ratios for the period before the increase in stream flow for Scotty Creek and Blackstone River, runoff ratios were infilled as a fraction of the runoff ratios at Jean Marie River and Blackstone River. For the available period of record, runoff ratios at Scotty Creek were smaller than the runoff ratios at Jean Marie River by a factor of 0.87 (*n*: 16; *SD*: 0.07), whereas runoff ratios at Blackstone River were smaller than the runoff ratios at Birch River by a factor of 0.98 (*n*: 21; *SD*: 0.14). The pairings of the two basins were chosen on the basis of each basin's proximity to the other (Figure 1). The low SDs suggest that it is valid to estimate runoff ratios for Scotty Creek and Blackstone River for the period before hydrometric data were available by extrapolating from the runoff ratios from Jean Marie River and Birch River, respectively, using the aforementioned factors.

After the appropriate ranges for runoff ratios were found for the period prior to the increase in stream flows (i.e. 1976–1995), the Kendall-Theil robust line test was used to calculate the amount of precipitation that would be expected to occur in 2012. That value was then multiplied by the runoff ratios to determine expected runoff rates in the period prior to the increase in stream flow, given increased levels of precipitation.

Permafrost-thaw-induced land-cover changes are hypothesised to increase both the area of hydrologically connected wetlands that contribute to stream flow and the area of the peat plateaus that provide primary runoff to the channel fen. The image analysis will quantify the change in the area (m^2) and edge length (m) of the area that contributes to stream flow (i.e. wetland areas that have a hydrological connection to the channel fen) within the 6 km^2 AOI. These calculated areas are multiplied by the expected runoff rates calculated earlier to provide estimates for the increases in runoff resulting from the direct increase in area of wetlands connected to the drainage network and the increases in the area of the peat plateaus that contribute primary runoff to the adjacent connected wetlands.

To estimate for the magnitude of the total runoff increase, the increases to streamflow-contributing areas, primary runoff contributing areas, amount of water released by the thawing permafrost bodies and additional inputs from increased winter flows were combined. The sum of these components were then compared with the observed increase calculated as the difference between the 1996 and 2012 annual runoff values on the Kendall-Theil robust line for each basin.

RESULTS AND DISCUSSION

Increases to stream flow

An increase in annual total runoff began in 1996 for all basins (Figure 5). The two basins with long-term records (Birch and Jean Marie Rivers) do not display any significant trends for the period prior to 1996; however, they display significant ($p \leq 0.01$) positive trends from 1996 to 2012 (annual increases of 8.3 and 9.4 mm/year, respectively). Scotty Creek ($p \leq 0.01$) and Blackstone River ($p \leq 0.05$) also show significant trends from 1996 to 2012 with annual increases of 8.5 and 6.6 mm/year, respectively. Precipitation patterns increased significantly ($p \leq 0.1$) over the period of 1996–2012 at a rate of 5.7 mm/year, however, did not show a significant change from 1976–1995 (Table II). It should be noted that not all of this precipitation contributes directly to stream flow (i.e. some is retained as storage and/or lost as ET). For the purposes of this paper, the precipitation that contributes to stream flow will hereinafter be referred to as 'effective precipitation'. Total runoff increased by 112 mm (Blackstone), 137 mm (Scotty), 141 mm (Birch) and 160 mm (Jean Marie) between the years 1996 and 2012 (Table II). Runoff ratios also increased significantly ($p \leq 0.1$) in all four basins since 1996 (Table II).

Bog capture

The thawing of permafrost is thought to change the proportion of land-cover types in wetland-dominated discontinuous permafrost environments. This is expected to have on impact on the movement and storage of water within these basins. An understanding of runoff generation processes coupled with the partitioning of moisture inputs is necessary to predict how these changes may affect the basin hydrograph. As thawing permafrost

Table II. Annual runoff, runoff ratios and baseflows for the four gauged basins in the lower Liard River valley, prior to and post apparent increase in stream flow (1996)

Basin	Annual runoff: 1976–1995			
	Average runoff (mm)	Total change (mm)	Average change/year (mm)	Average change/year (%)
Birch	127.8	−22.9	−1.1	−0.9
Jean Marie	95.5	−12.4	−0.6	−0.7

Basin	Runoff ratios: 1996–2012			
	Average runoff ratio	Total change	Average change/year	Average change/year (%)
Birch	0.400	***0.188***	***0.011***	***4.2***
Jean Marie	0.327	***0.236***	***0.014***	***7.0***
Blackstone	0.385	*0.200*	*0.012*	*4.0*
Scotty	0.278	***0.218***	***0.014***	***8.3***

Basin	Annual runoff: 1996–2012				Annual baseflow: 1996–2012				
	Average runoff (mm)	Total change (mm)	Average change/year (mm)	Average change/year (%)	Average baseflow (mm)	Total change (mm)	Average change/year (mm)	Average change/year (%)	2012 Baseflow contribution (%)
Birch	216.0	***140.8***	***8.3***	***6.4***	5.1	***6.7***	***0.4***	***38.8***	***2.9***
Jean Marie	177.0	***160.1***	***9.4***	***11.4***	11.4	***6.8***	***0.4***	***5.0***	***6.1***
Blackstone	207.3	***111.5***	***6.6***	***4.1***	5.6	***5.5***	***0.3***	***13.1***	***2.9***
Scotty	149.0	***136.8***	***8.5***	***10.9***	1.4	*0.9*	*0.1*	*18.5*	*0.6*

Period	Fort Simpson annual precipitation			
	Average precipitation (mm)	Total change (mm)	Average change/year (mm)	Average change/year (%)
1976–1995	424.7	−43.0	−2.3	−0.5
1996–2012	539.7	*97.2*	*5.7*	*1.1*

Precipitation values were obtained at Fort Simpson, NWT, over the period of 1976–2012. Mean change (% and mm) are derived from the Kendall-Theil robust lines. Italicised, bolded and underlined values represent statistical significance at $p \leq 0.1$, $p \leq 0.05$ and $p \leq 0.01$, respectively.

causes wetlands to coalesce, it appears that storage capacity in bogs is decreasing, whereas areas contributing to flow are increasing. There is potential for the runoff contributing area to increase significantly when thaw removes a permafrost plateau from between a channel fen and a formerly isolated bog. This process is depicted in Figure 6(c) and (d) and is termed 'bog capture'.

Analysis of remotely sensed imagery and aerial photographs shows that permafrost covered $3.27 \pm 0.3\,km^2$ in 1977 and decreased to $2.85 \pm 0.23\,km^2$ by 2008 over the $6\,km^2$ AOI (Table III). This represents a 13% (±2.7%) loss of permafrost coverage between 1977 and 2010. As described earlier, permafrost loss transforms forested area to wetland, and because these two cover types have characteristic hydrological functions, this loss produces local changes to the hydrological cycle. The expansion showed that permafrost-free terrain reduced the proportion of the landscape that is hydrologically isolated from the basin drainage network.

In 1977, the drainage network covered 31.4% of the $6\,km^2$ AOI. The drainage network expanded to cover 41.8% of the AOI by 2010. It is worth noting that the resultant increase of the drainage network (10.4%) is greater than the overall increase in permafrost-free areas (7.0%). This illustrates that growth of the drainage network is exhibiting a nonlinear response to warming. As outlined in Figure 4, exclusive of ET, precipitation falling directly on the channel fen or on the flanks of plateaus adjacent to the channel fen will be conveyed to the basin outlet. The effects of ET require more study; however, the relatively high ET rates calculated by Wright *et al*. (2008) on saturated surfaces (i.e. wetlands) may offset some of the effects of precipitation inputs.

The area of the drainage network increased by a factor of 1.33, resulting in an increase in runoff of 37–61 mm, the highest of all factors contributing to increases in runoff. Table IV shows the depth of runoff that would be

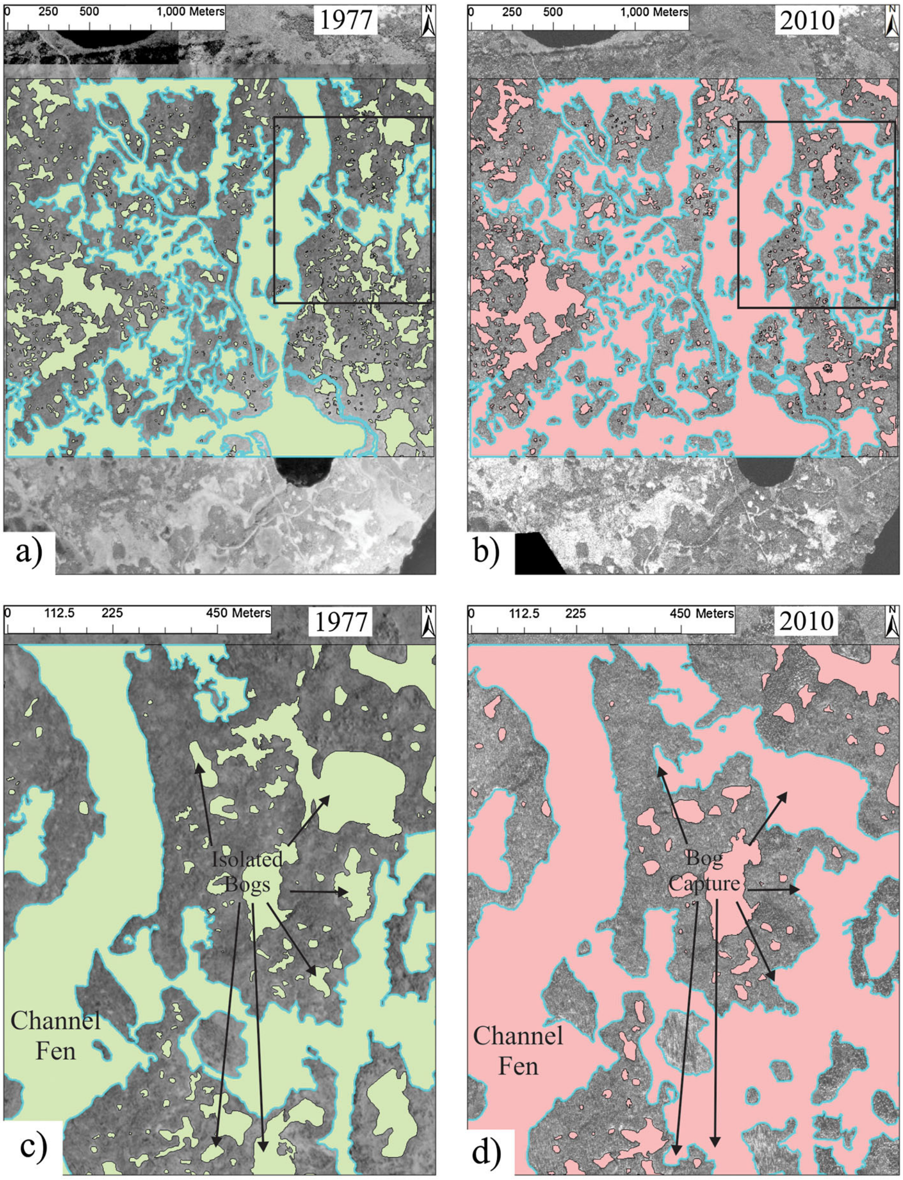

Figure 6. A 6 km^2 area of interest at Scotty Creek basin for (a) 1977 and (b) 2010 outlining permafrost and permafrost-free areas. Permafrost-free areas (i.e. wetlands) are shaded in green (1977) and red (2010). Sections of wetland outlined in blue represent the area that is connected to the basin drainage network. (c) 1977 and (d) 2010 are insets highlighting the concept of bog capture

Table III. Changes in permafrost coverage over a 6 km^2 area of interest at Scotty Creek, NWT

Year	Permafrost (m^2)	Permafrost (%)	Drainage contributing area (m^2)	Drainage contributing area (%)	Drainage perimeter (m)	Isolated bogs	Isolated area (m^2)	Isolated area (%)
1977	3 274 578	54.6	1 884 661	31.4	73 862	553	840 761	14
2010	2 847 458	47.5	2 508 154	41.8	86 754	404	644 388	11
Change	427 120	7.1	623 493	10.4	12 892	149	196 373	3.3

expected in each basin given the increase in drainage area expected from the image analysis at Scotty Creek.

Of the four basins that were analysed, Scotty Creek and Jean Marie River displayed the highest rates of increased stream flow, whereas Birch and Blackstone were noticeably lower. Using conservative estimates of runoff ratios (i.e. the lowest plausible runoff ratio), the expected runoff in the Birch and Blackstone River basins exceeds the expected

Table IV. Expected runoff values for the four basins in the lower Liard River valley calculated with the 25% trimmed mean runoff ratios from 1976 to 1995 (beginning of period of apparent increase in stream flow) with additions from the conversion of ground ice to water as a result of permafrost thaw, increased baseflow contributions, increased drainage areas and increases to plateau runoff ('new total')

				Additional annual inputs (mm)			
Basin	2012 Precipitation with 1976–1995 runoff ratios (mm)	+	−	Channel precipitation	Primary runoff contributing areas	Thawed permafrost	Additional baseflow
Birch	183	50	47	60.5	31.9	9	6.7
Blackstone	174	38	43	57.5	30.3	9	5.5
Scotty	113	19	25	37.3	19.7	9	0.9
Jean Marie	132	22	29	43.6	23.0	9	6.8
Basin	New total (mm)	+	−	Observed 2012 runoff (mm)	Mean unaccounted Water	Range of unaccounted water (mm)	
Birch	290.5	50	47	270.8	−19.7	−69.3	27.6
Blackstone	275.7	38	43	273.3	−2.4	−39.9	40.7
Scotty	179.5	19	25	215.3	35.9	17.0	61.1
Jean Marie	213.9	22	29	243.0	29.1	7.0	58.5

Plus/minus columns indicate the range of runoff that may be expected (conservative and non-conservative estimates of runoff ratios). Unaccounted water is calculated by subtracting the new total from the expected 2012 runoff (derived from the Kendall-Theil robust line).

runoff after accounting for additional inputs (Table IV). Quinton *et al.* (2003) show that the per cent cover of flat bog at Scotty Creek is 10.2% compared with 3.4% (Table I) at Blackstone. This suggests that the expansion rate of the drainage area calculated at Scotty Creek may not be transferrable to other basins that have different proportions of bogs to fens. The number of isolated bogs in the AOI at Scotty Creek decreased from 553 to 404 and decreased in area by 3.3% over the 33-year study period. If the original number of bogs in a basin is low (i.e. Blackstone), the potential for bog capture resulting from permafrost thaw would also be low.

As a basin's storage capacity decreases (i.e. decreases in isolated bog coverage), a concomitant increase occurs in the area contributing to basin runoff. As such, the implications of a once-isolated bog forming a connection with the channel fen are twofold: (1) The area contributing to basin storage decreases and (2) the basin drainage density increases. As the area contributing to streamflow expands and previously isolated bogs are becoming fully or seasonally connected, they are also contributing their catchment area to the basin drainage network. Accordingly, bogs with large catchment areas that are located near the basin drainage network but are still hydrologically isolated have the potential to discharge large quantities of water that were previously held as storage given further permafrost thaw. Figure 7 shows that the average annual increases in basin runoff (%) and basin runoff ratio (%) are positively correlated with the per cent cover of flat bog in a basin (R^2: 0.78 and 0.81; n: 4) and negatively correlated with the per cent cover of fen (R^2: 0.67 and 0.88; n: 4) and drainage density (R^2: 0.83 and 1.00; n: 4) (where R^2 represents the coefficient of determination and n represents the number of basins). Therefore, it appears that basins with a higher ratio of flat bogs to channel fens and relatively low drainage densities may be the most vulnerable to rising stream flows given further permafrost thaw.

Changes in runoff contributing area

Growth of the secondary plateau contributing area is difficult to quantify from aerial photographs and remotely sensed imagery because of uncertainties of drainage divides on peat plateaus. Because of the difficulties in quantifying the area on the plateau contributing to runoff and the flux of water being added, additions from the growth of the secondary runoff contributing areas (Figure 4) were not included in this study. The expansions of these areas and the resultant increases in runoff have yet to be quantified and are the objectives of future studies.

The primary contributing areas (i.e. edge of peat plateaus) maintain a relatively equal length and slope throughout the basin. Chasmer *et al.* (in review) measured the length and slopes of 15 edges using remotely sensed imagery. They found a mean edge length of 17 m (*SD*: 7.3 m) with a mean slope of 0.041 (*SD*: 0.019). Using a hand-held clinometer, Quinton and Baltzer (2013) measured the slopes of nine

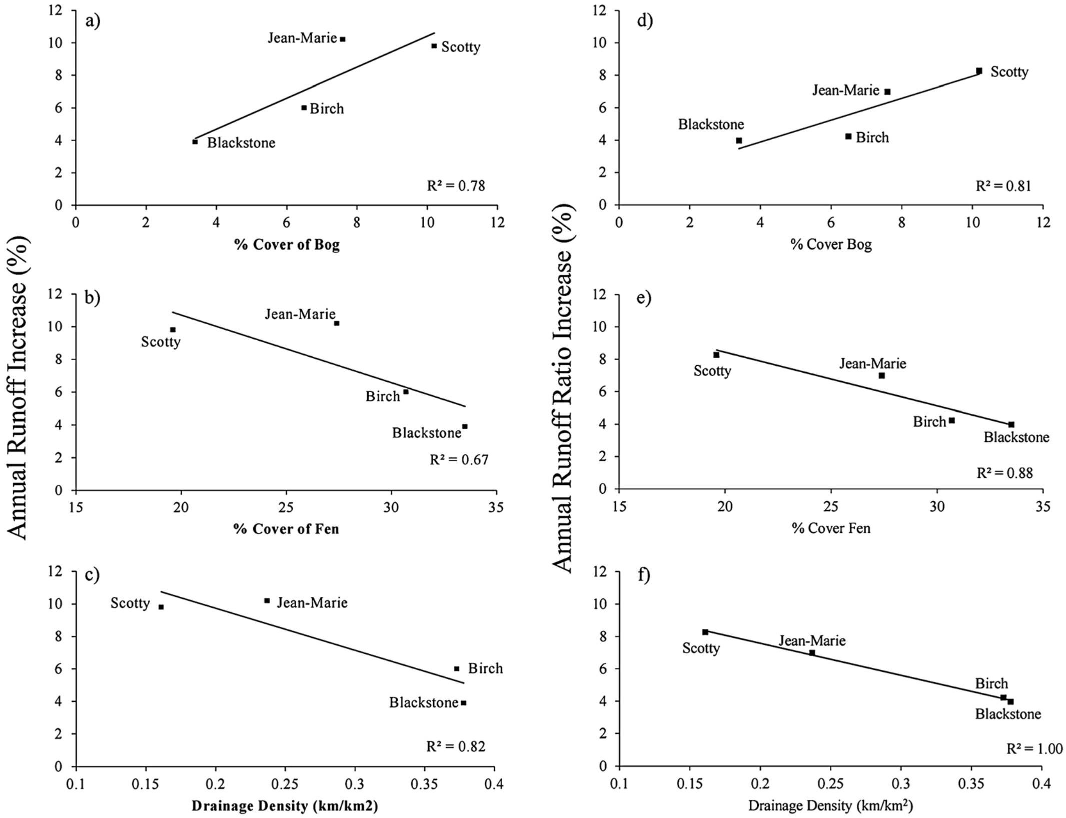

Figure 7. Annual increase (%) in runoff and runoff ratios plotted against: (a,d) per cent cover of bog, (b,e) per cent cover of fen and (c,f) drainage density for four basins in the lower Liard River valley

edges and found a mean of 0.041 (*SD*: 0.006). This relative uniformity in the length and slope of edge areas can be useful in predicting increases in plateau runoff resulting from expansion of the drainage network. As bogs expand and amalgamate with neighbouring bogs, their surface area increases, their shapes become more complex and the length of their boundaries with adjacent plateaus increases. As the edge length of wetlands increases, so too does the area in the adjacent plateaus that contributes runoff to them. This growth of the primary runoff contributing area is represented by a simple area equation ($A = l \times w$), where if width is relatively constant (as is shown by the uniformity of slope lengths and angles), the area of the plateau contributing to primary runoff should be increasing given an increase in edge length. Therefore, it is hypothesised that the total area on the plateau that contributes subsurface flow to the fen would increase linearly by the same amount as the total edge length of the drainage network. This is thought to be one method of quantifying the growth of primary runoff contributing areas; however, it is recognised that further study is needed to increase certainty in the growth of these runoff contributing areas. The edge length of the drainage network grew by a factor of 1.18 between 1977 and 2010, resulting in an additional input of 20–32 mm of runoff (Table IV).

Changes to precipitation patterns

Effective precipitation increased by between 18 and 30 mm in the four basins between the years of 1996 and 2012. Although these additional precipitation inputs do not entirely account for the observed increase in stream flows, changes in temporal precipitation patterns may also have an effect on basin runoff. Atmospheric moisture inputs have been increasing (Table II) in the Fort Simpson region, but changes to the timing, magnitude and intensity of precipitation events can also alter the basin hydrograph in subarctic environments (Spence *et al.*, 2011). There are three distinct zones of horizontal saturated hydraulic conductivity (K_s) on a peat plateau (Quinton *et al.*, 2008). The top 0.1 m is a zone of uniformly high K_s, followed by a transition zone, with depths below 0.2 m being a zone of uniformly low K_s. Precipitation events that occur when the water table occupies the zone of uniformly high K_s will produce greater runoff. During the spring period, the frost table and the water table perched above it occupy this zone of uniformly high K_s. As the thaw season progresses and the active layer thickens, the frost table and water table lower into zones of decreasing K_s (Quinton *et al.*, 2008). Because of the relatively high drainable porosity of peat at depth, late-season precipitation events that are suitable to raise the water table back into the zone of high K_s can also produce greater runoff. Therefore, not only climate-induced variation in annual precipitation but also the timing, magnitude and intensity of the precipitation may lead to significant changes in mean runoff. A detailed analysis of the effect of changing temporal precipitation patterns on total basin runoff in the lower Liard River valley is beyond the scope of this paper and requires further investigation.

General circulation models predict an increase in precipitation at higher latitudes under a changing climate (Manabe *et al.*, 2004; Wu *et al.*, 2005). Prior to the apparent increase in stream flow, annual runoff was significantly correlated to annual precipitation at both gauged river basins (Birch: R^2: 0.45; p: 0.002; Jean Marie: R^2: 0.42; p: 0.003). Plotting runoff against precipitation for the period of 1996–2012 yields R^2 values between 0.13 and 0.17 with non-significant p-values for the four gauged basins, indicating that precipitation has been exerting less control on runoff than it has in the past (Figure 8). Precipitation depth at Fort Simpson has been increasing; however, this increase does not explain the observed increase in runoff from the four basins. Instead, it appears that antecedent soil moisture conditions are a better predictor of runoff, as it governs the thickness and position of the saturated layer and therefore K_s of this layer, as well as the soil storage capacity for future precipitation events. Long-term ET data are not available for the Scotty Creek basin, and therefore, the changes to ET rates could not be calculated. It should be noted that Wright *et al.* (2008) calculated that the ET loss in bogs (i.e. from the saturated surface) was nearly double that of an adjacent peat plateau (with an unsaturated surface). Therefore, it appears that as permafrost thaw converts peat plateaus to wetlands, the ET flux may increase and possibly dampen the response of increased stream flows.

It is worth noting that this period of apparent increasing stream flows also included the largest (2008) and the second (2012), third (1999), fourth (2007) and sixth (2005) largest recorded total annual snowfalls for the 1898–2012 period of record (Table V). There is a possibility that the apparent increase in stream flow was driven in part by these several large annual snowfalls

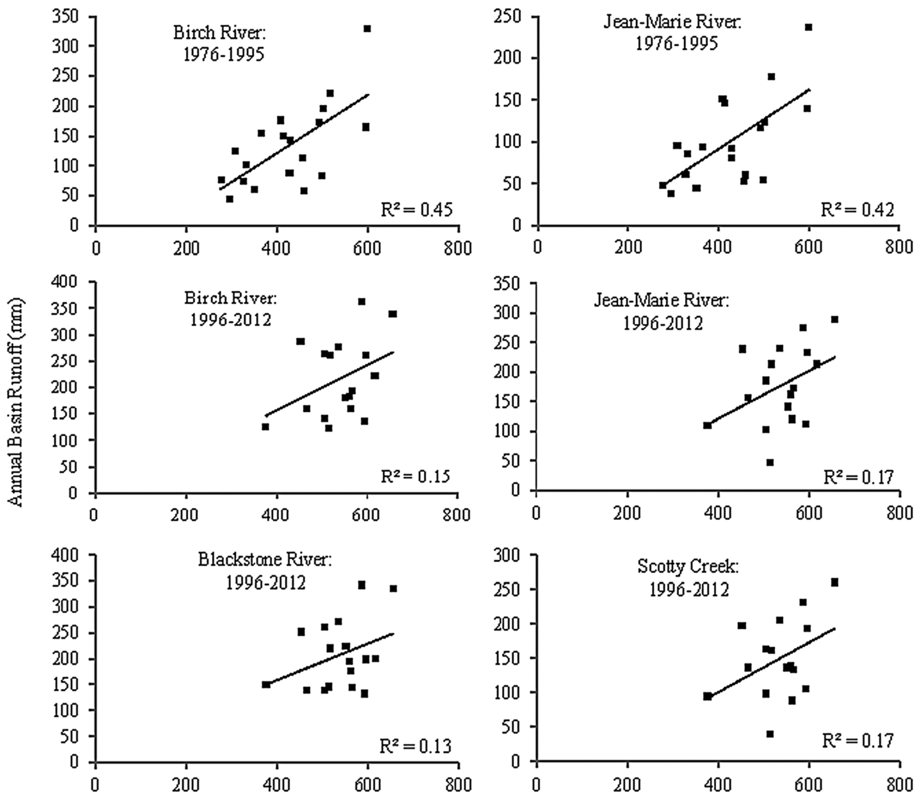

Figure 8. Plots of annual basin runoff against annual precipitation for two time periods (1976–1995 and 1996–2012) for four basins in the lower Liard River valley. For the period of 1976–1995, annual basin runoff is significantly correlated to annual precipitation for both Birch River and Jean Marie River ($p < 0.01$)

Table V. Ten highest years for corrected precipitation (total, snow and rain) data from 1898 to 2012 (Mekis and Vincent, 2011) from the Environment Canada station, Fort Simpson, NWT

Rank	Year	Total (mm)	Year	Snow (mm)	Year	Rain (mm)
1	2009	656.2	2008	341.5	1988	399.4
2	2012	616.9	2012	331.1	2009	378.0
3	1988	600.1	1999	318.6	1982	365.9
4	2011	597.9	2007	307.7	2010	350.1
5	1982	597.5	1983	301.3	1997	345.4
6	1999	594.9	2005	295.1	2011	331.5
7	2010	589.2	1917	294.2	2001	327.5
8	2008	566.1	2006	289.8	1996	319.9
9	1997	563.9	1972	279.4	1981	309.3
10	2000	560.1	2009	278.2	1990	306.0

since 1999. Studies at Scotty Creek by Hayashi *et al.* (2004) provided some insights into possible effects on stream discharge of inter-annual variation of basin snow storage. Their study examined the isotopic and chemical signatures at several points along the basin drainage network including the basin outlet and found that the direct snowmelt contribution was less than half of total basin discharge, indicating the importance of the water stored over winter in the interconnected bogs and channel fens. Carey *et al.* (2012) observed similar results in a wetland-dominated discontinuous permafrost basin, where snowmelt water during the freshet accounted for between just 10% and 26% of total runoff. It is possible that the effect of large snow storage on basin discharge may carry forward into the following year, and several large snow years in close succession may promote a period of increasing stream flow, because of elevated water storage levels in the bogs and channel fens. Other studies have linked increased precipitation levels with rising stream flows in Arctic and subarctic basins (Peterson *et al.*, 2002; Déry *et al.*, 2009); however, few have studied the role of changing soil moisture conditions as a result of increased snowfall at the headwater catchment scale. The concept of snowmelt contributing to higher water storage levels may explain why runoff has been less dependent on total annual precipitation in recent years (Figure 8).

Moisture inputs from permafrost thaw

Additional moisture inputs resulting from the conversion of ground ice to water from permafrost thaw are estimated to be approximately 7–9 mm/year (Table VI). As initial ice content is lowered, the amount of additional moisture input decreases by just over 0.1 mm/year (Table VI). This amount is minimal over the short term; however, it may be of more importance over longer periods as antecedent ice content continues to decline. It should be noted that the values calculated here were obtained using average thaw rates over the past 33 years. Recent trends of more rapid permafrost thaw (Lantz and Kokelj, 2008; Quinton *et al.*, 2011) indicate that this may be a conservative estimate of annual moisture inputs from this thawing permafrost. For this reason, we used the uppermost estimate of 9 mm in calculations for Table IV. These values are representative of field conditions at Scotty Creek and will vary from basin to basin depending on differing permafrost coverage in other basins (both vertical and horizontal extents). The findings that direct moisture inputs from thawing permafrost are only a secondary contribution to increasing stream flows that are consistent with McClelland *et al.* (2004) who calculated that about 4 m of vertical permafrost thaw would be required to achieve the observed increase in stream flow of several large rivers in permafrost-dominated basins in Eurasia. It appears that additional moisture inputs from the conversion of ground ice to water are likely a secondary factor.

Additional groundwater contributions

The literature frequently suggests that permafrost thaw is reactivating groundwater pathways and thereby increasing baseflow and subsequently stream flow rates (St. Jacques and Sauchyn, 2009; Walvoord and Striegl, 2007). Using winter flows as a proxy to estimate changes to baseflow, it is shown that baseflow contributions appear to be minimal (≤7% of total annual runoff); however, each basin shows a

Table VI. The annual additional moisture inputs from the conversion of ice to water resulting from permafrost thaw for selected years

Year	Area Covered by permafrost (%)	Area covered by permafrost (m^2)	Area of permafrost lost (m^2)	Volume of ice over basin (m^3)	Annual additional moisture inputs (mm)
1977	54.6	82 992 000	179 062	1 432 492	8.6
1980	54.0	82 008 145	176 939	1 415 510	8.5
1990	51.8	78 728 630	169 863	1 358 904	8.1
2000	49.6	75 449 115	162 787	1 302 297	7.8
2010	47.5	72 169 600	155 711	1 245 691	7.5

Permafrost depth and porosity are held constant at 10 m and 0.8, respectively. Area coverage is based on the Scotty Creek basin.

significant increase in baseflow during the period of 1996–2012 ($p \leq 0.1$ for Scotty; $p \leq 0.05$ for Blackstone; and $p \leq 0.01$ for Birch and Jean Marie). Baseflow has increased by 0.9 (Scotty) to 6.8 (Jean Marie) mm/year in the four basins. Annual baseflow contributions to total annual stream flow range from 0.6% (Scotty) to 6.1% (Jean Marie) on the basis of estimates in 2012 using the Kendall-Theil robust line. Even if it is assumed that increases to baseflow can be approximated by using winter flow hydrometric data (which is already a conservative assumption), these results show that the reactivation of groundwater pathways resulting from permafrost thaw is not sufficient to support the observed increases to stream flows in the lower Liard River valley and appears to be a secondary contributor.

CONCLUSIONS

Stream flow has increased in the lower Liard River valley in the NWT as a result of climatic drivers. Although precipitation has increased in the region, this does not entirely explain the observed increases in stream flow. The current paradigm in the literature suggests that the opening of groundwater pathways resulting from permafrost thaw is a primary driver behind increasing stream flows. We show that in the lower Liard River valley, increasing groundwater inputs are a relatively small component. Here, we have outlined a number of alternative mechanisms, induced by land-cover change via long-term permafrost thaw, which may better explain the magnitude of these increases. These land-cover changes have the potential to fundamentally alter the hydrology of a basin as the relationship between bogs and peat plateaus changes with ongoing permafrost thaw. Bogs that were formerly isolated are now becoming incorporated into the basin drainage network. This is shown by a strong correlation between increased stream flow and the per cent cover of bogs in a basin, indicating potential areas of vulnerability. As permafrost thaws, surficial pathways for water become more direct as runoff-generating land-cover types coalesce. Increases to the size of the basin drainage network allow for more channel precipitation and increased runoff from adjacent plateaus. We have shown that surface pathways have opened up at a rate disproportionate to permafrost thaw and can therefore be viewed as a nonlinear response to warming.

Further work is necessary to quantify the additional inputs from primary and secondary contributing areas. A conceptual model describing the partitioning of runoff water from peat plateaus elucidates the need for future studies to develop a better understanding of secondary runoff and seasonal bog interconnection. Permafrost thaw and associated land-cover changes appear to be a primary driver in changing the basin water balance, and further work is necessary to fully understand this implication on northern water resources.

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