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The hydrology of interconnected bog complexes in discontinuous permafrost terrains

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Abstract:

In the zone of discontinuous permafrost, the cycling and storage of water within and between wetlands is poorly understood. The presence of intermittent permafrost bodies tends to impede and re-direct the flow of water. In this region, the landscape is characterized by forested peat plateaus that are underlain by permafrost and are interspersed by permafrost-free wetlands. These include channel fens which convey water to the basin outlet through wide, hydraulically rough channels and flat bogs which are typically thought to retain moisture inputs as storage. Field studies conducted at a peatland-dominated landscape near Fort Simpson, Northwest Territories, Canada, indicate the presence of ephemeral drainage channels that form a cascade of connected bogs that ultimately discharges into a channel fen. Consequently, understanding bogs as dynamic transmitters of surface and subsurface flows, rather than simple storage regions, calls for further examination. Whether bogs act as either storage features or flow through features has a direct impact on the runoff contributing area in a basin. Here, two adjacent series of bog cascades were gauged over two consecutive years to determine spatial and temporal changes in effective runoff contributing areas. It was found that runoff varies significantly between two adjacent bog cascades with one cascade producing 125 mm of runoff over the 2-year period, while the other yielded only 25 mm. The bog cascades are primarily active during the snowmelt season when moisture conditions are high; however flows can also be generated in response to large rain events. It is proposed that bog cascades operate under an 'element threshold concept' whereby in order for water to be transmitted through a bog, the depression storage capacity of that bog must first be satisfied. Our work indicates that whether bogs act as storage features or flow-through features has a direct impact on the runoff contributing area in a basin. Neglecting to represent connected bogs as dynamic transmission features in the landscape is shown to underestimate water available for streamflow by between 5 and 15%, and these systems are therefore a key component of the water balance in discontinuous permafrost regions.

KEY WORDS peatlands; bogs; hydrology; discontinuous permafrost; wetlands; runoff

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INTRODUCTION

Climate warming in the discontinuous permafrost region of northwestern Canada is occurring at one of the fastest rates on earth and is threatening to increase mean annual air temperatures above zero degrees (Johannessen *et al*., 2004; Kirtman *et al*., 2013). One of the most prominent impacts of this warming is the degradation of permafrost (i.e. ground that is below 0 °C for two or more consecutive years) in this region. Permafrost thaw is prevalent in North America, especially along the southern margin of discontinuous permafrost (Kwong and Gan, 1994; Jorgenson and Osterkamp, 2005). Thawing permafrost has major implications on landscapes, including, but not limited to: (1) ground subsidence (i.e. thermokarst development) (Gooseff *et al*., 2009); (2) changes in ecosystem composition, structure and function, and disturbance frequency and severity (Beck *et al*., 2011; Grosse *et al*., 2011); and (3) changes to the hydrological regime including changes to the timing and magnitude of runoff (Quinton and Baltzer, 2014; Jones and Rinehart, 2010; St Jacques and Sauchyn, 2009). In respect to the latter, the presence of permafrost typically impedes the transmission and movement of subsurface water (O'Donnell *et al*., 2011), and thereby exerts a significant control on the cycling and storage of water within a basin. In the lower Liard River valley of the Northwest Territories (NWT), Canada (Figure 1), this southern edge of permafrost underlies the northern boreal forest where there is extensive coverage of peatlands (Aylesworth and Kettles, 2000). Forests situated on thawing ice-rich permafrost bodies, such as those in organic terrains with high porosity (i.e. peat plateaus or palsas), are at risk of conversion to wetland if the permafrost body thaws (Osterkamp *et al*., 2000; Jorgenson

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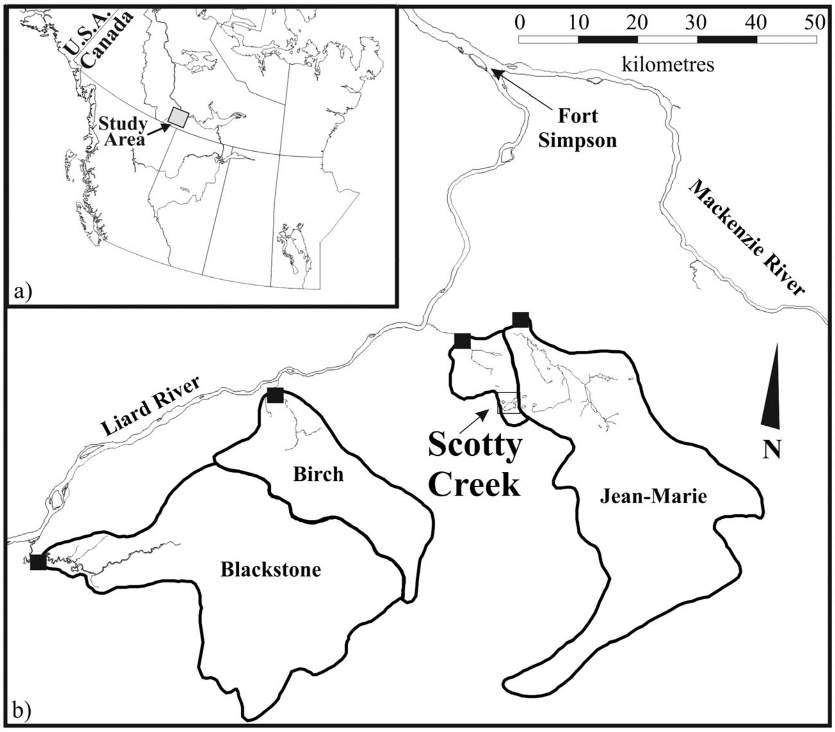

Figure 1. a) Location of the lower Liard River valley in the Northwest Territories, Canada; b) inset of the Scotty Creek Research Basin

et al., 2001; Quinton *et al*., 2011). Collapse scar bogs and fens are common thermokarst features in the northern boreal forest, and can result from even minor changes to the energy balance on a forested peat plateau (Quinton *et al*., 2009; Baltzer *et al*., 2014).

As the proportion of wetlands increases over the landscape, there is also an increase in the hydrological connectivity of these features (Connon *et al*., 2014). In the headwaters of the lower Liard River valley, flow was found to be largely restricted to channel fens (Hayashi *et al*., 2004), features that are bordered by permafrost bodies and route water to the basin outlet. Flat bogs are internal collapse features on peat plateaus and are devoid of permafrost. The water stored in these bogs is surrounded on all sides by raised permafrost (Jorgenson and Osterkamp, 2005), typically serving to contain the water within the bog. In some cases, ephemeral channels have been found to cut through permafrost bodies and provide a hydrological connection between bogs (Hayashi *et al*., 2004). The amount of water transmitted through these channels and the subsequent effect on the basin hydrograph has yet to be quantified. It is likely that as permafrost continues to thaw, the degree of connectivity between bogs will continue to increase and this mechanism of runoff transmission will play an increasingly important role.

Our current understanding of drainage channels connecting bogs in discontinuous permafrost environments is based on few investigations. To our knowledge, the first recorded observation of drainage channels connected to collapse scar bogs was by Thie (1974) at a subarctic site in central Manitoba. The author commented on the influence that these channels may have on thawing permafrost, but did not discuss their hydrological implications. Vitt *et al*. (1994) also mention the presence of drainage channels connected to bogs in northern Alberta, but like Thie (1974), these authors touch only on their importance to permafrost thaw in the adjacent bogs. Vitt *et al*. (1994) analysed the stratigraphy in the peat cores in these connected bogs and found that unlike other collapse scar features in the region, these bogs did not show evidence of previously drier conditions, suggesting that permafrost may never have been present in these features and that permafrost developed around them. This indicates that in some regions, drainage channels between bogs may not be an artifact of permafrost thaw, but instead, an impedance to the aggradation of permafrost as energy from connected bogs may be advected through these features.

More recently, Connon *et al*. (2014) presented a conceptual model for the potential partitioning of runoff water in a wetland-dominated discontinuous permafrost basin, and briefly discussed the potential influence of ephemeral drainage channels between bogs. In their paper the authors show that runoff from peat plateaus can either: (1) flow directly into the channel fen (i.e. primary runoff); (2) flow through a series of ephemerally connected and cascading bogs eventually discharging into a channel fen (i.e. secondary runoff); or (3) flow into an isolated or temporarily disconnected bog where water is retained as storage and is only lost through evapotranspiration. The authors provide evidence showing that the primary runoff contributing areas are increasing as a result of permafrost thaw; however the effects of secondary runoff were left unquantified. Secondary runoff contributing areas are ephemeral in nature and thought to be dependent on the depression storage capacity of the bogs. Buttle *et al.* (2012) provide a review of temporary (i.e. intermittent, ephemeral and episodic) stream hydrology in Canada and note that there is a poor understanding of the processes that govern ephemeral systems in the headwaters of permafrost basins. The authors discuss that temporary stream dynamics are indicative of the behaviour of hydrologic connectivity within a basin; this is analogous to the ephemeral bog connections studied here. Runoff generation is dependent on the fractional area contributing to flow; therefore to properly understand the hydrology of peat plateau–bog complexes it is imperative to understand how the runoff generating area changes with time and space. A critical step towards this is to develop an understanding of how secondary runoff contributing areas function and how and when they contribute to flow.

We hypothesize that secondary runoff contributing areas operate under different degrees of connectivity, dependent on existing storage levels in the bogs. In a fully connected system, such as occurs during large snowmelt or precipitation events, water is transmitted through all bogs in a bog cascade. As water levels drop and the channels cease to transmit water, the total contributing area to the terminal bog in a cascade shrinks. The depression storage capacity, unique to each bog in the cascade, must be exceeded in order for flow to resume. The goal of this paper is to investigate the mechanism(s) controlling the transmission of water through bog cascades and to quantify the amount of runoff produced from these systems. Supported by this investigation, we will: (1) demonstrate that connected bogs can generate non-negligible flows and quantify the magnitude of their influence on the basin water balance; (2) characterize the processes that control runoff generation in these systems at a well monitored field site; and (3) identify challenges to predicting runoff from secondary contributing areas.

STUDY SITE

Field studies were conducted in the headwaters of the Scotty Creek Research Basin (SCRB), a 152-km^2 watershed located about 50 km south of Fort Simpson, NWT (Figure 1). The SCRB is typical of basins in the lower Liard River valley and consists of organic rich terrain underlain by discontinuous (~40%) permafrost (Quinton *et al.*, 2011). The basin has very little relief with an average slope of 0.0032 m m^{-1} (Quinton *et al.*, 2003). Mean annual air temperatures are $-3\,°C$, and average precipitation is 388 mm yr^{-1} with 62% falling as rain and 38% as snow (MSC, 2014). The region has short, warm summers (average July temperature of 17.4 °C) and long, cold winters (average January temperature of $-24.2\,°C$). The headwaters of the SCRB consist of four main land cover features: peat plateaus (43.0%), flat bogs (26.7%), channel fens (21.0%) and lakes (9.3%) (Quinton *et al.*, 2009). Permafrost exists exclusively under peat plateaus and is preserved by the large thermal offset created by an insulating layer of unsaturated peat. Peat plateaus are mainly populated by black spruce (*Picea mariana*) trees, whose root system is maintained by a dry vadose zone above the permafrost, as well as Labrador tea (*Rhododendron groenlandicum*) and ground lichen (*Cladonia* spp.). The peat plateaus are raised about 1 – 2 m above the adjacent wetlands owing to the volumetric expansion of their permafrost base. As a result, peat plateaus generate runoff into the flat bogs and channel fens (Wright *et al.*, 2009). Channel fens are wide (50 – 100 m), hydraulically rough features that convey water to the basin outlet. The fens have a floating vegetative mat consisting of various sedges (i.e. *Trichophorum alpinum*, *Eriophorum* spp., *Carex* spp.) and scattered tamarack (*Larix laricina*) (Garon-Lebrecque *et al.*, in revison). Flat bogs (sometimes referred to as collapse scar bogs) form on a peat plateau after disturbance (i.e. lightning, fire or anthropogenic causes) removes the tree canopy and alters the energy balance enough to induce thawing of the underlying permafrost (Robinson and Moore, 2000; Baltzer *et al.*, 2014). These features are dominated by *Sphagnum* mosses (i.e. *Sphagnum fuscum*, *S. angustifolium* and *S. riparium*) (Zoltai, 1993). An individual peat plateau may have a number of flat bogs within it; this system is referred to as a peat plateau–bog complex. Water stored in flat bogs is typically bound by the relatively impermeable permafrost of peat plateaus. As permafrost thaws, the degree of connectivity between flat bogs and channel fens is increasing (Quinton *et al.*, 2011; Connon *et al.*, 2014).

It has been observed that bogs can be hydrologically connected to the channel fen system in one of two ways: through an open, diffuse connection where moisture can interact between the two wetland systems year round (Quinton *et al.*, 2003); or through a series of connected bogs, forming a cascade along a very low topographic gradient with channels that cut through the permafrost plateau. Hereafter, the former are referred to as *open bogs* while the latter are referred to as *cascade bogs*. Isolated bogs are bogs in which there is no flow path to the channel fen. A conceptual model of this system is presented in Figure 2, illustrating the different flow paths which are dependent on moisture conditions at the time of

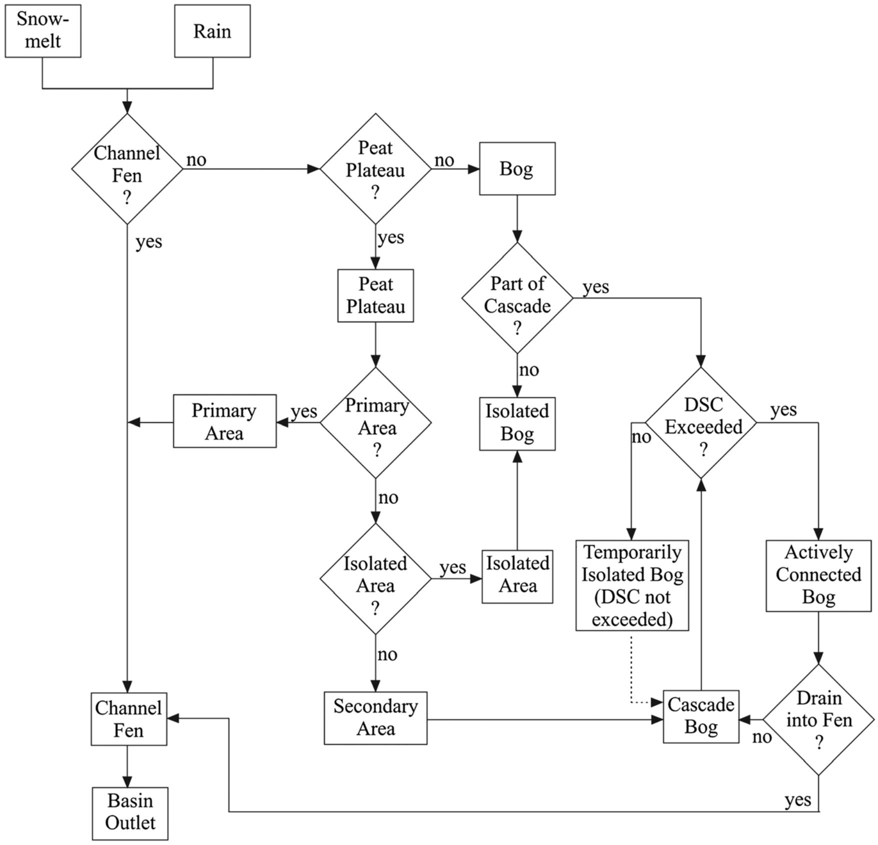

Figure 2. Conceptual diagram of the partitioning of precipitation inputs. The rectangles represent processes while diamonds represent questions. DSC: Depression Storage Capacity. Note that temporarily isolated bogs may reconnect with the rest of the cascade during rain events. Neglecting to include secondary runoff contributing areas (i.e. right-hand side of the diagram) as a source of input into the channel fen may underestimate basin stream flow

the input. Figure 3 shows the two series of monitored bog cascades at the SCRB as well as isolated bogs which do not have drainage to the channel fen. The two cascades are termed the west cascade and the east cascade. Each bog within the cascade has a unique number, and the numbers increase downstream (i.e. the most upstream bog in the west cascade is bog W-1, the most downstream is bog W-6). The drainage channels (or connections) that connect the bogs are labelled according to the bog upstream of it (i.e. the drainage channel after bog W-6 is termed 'connection W-6'). It should be noted that the west cascade has two terminal drainage channels, one at the northwest corner of bog W-3 (outlet is called W-3A) and one flowing out of bog W-6. Historical photographs indicate that bog W-3 is the amalgamation of two previously separate bogs and this explains the second outlet. Total discharge from the west cascade includes the sum of both outlets unless otherwise indicated.

Bog cascades

Bog cascades are formed on a peat plateau–bog complex when drainage channels link two or more flat bogs and allow for the transmission of water between bogs driven by typically very small hydraulic gradients (i.e. 0.001 – 0.002). Every bog on a peat plateau has an associated 'bogshed'. The bogshed is the area of the plateau that contributes runoff to that bog. These bogshed boundaries are very difficult to identify from light detection and ranging (LiDAR) imagery or field surveys because of flat topography; therefore bogsheds were approximated using a nearest-neighbour approach whereby each point on the peat plateau is presumed to drain to the closest bog. Ground-truthing (using a differential global positioning system) indicates that this is reasonable as a first-order approximation. Inputs into the two bog cascades are presumed to be strictly meteoric because of the presence of permafrost surrounding the bogs and clay-rich glacial till with low hydraulic conductivity [$K_S = 1 \times 10^{-9}$ – $1 \times 10^{-10}\,\mathrm{m\,s^{-1}}$ (Hayashi *et al.*, 1998)] underlying the system. The transmission of water within the bog cascades is restricted to drainage channels cutting through the peat plateau; however the east cascade may have diffuse supra-permafrost groundwater flows between bog E-3 and E-5 (Figure 3). The average cross-sectional area of surface water in the channels during the spring freshet is $1.0\,\mathrm{m}^2$ (n: 12; standard deviation: 0.5) with a range of $0.4\,\mathrm{m}^2$ to

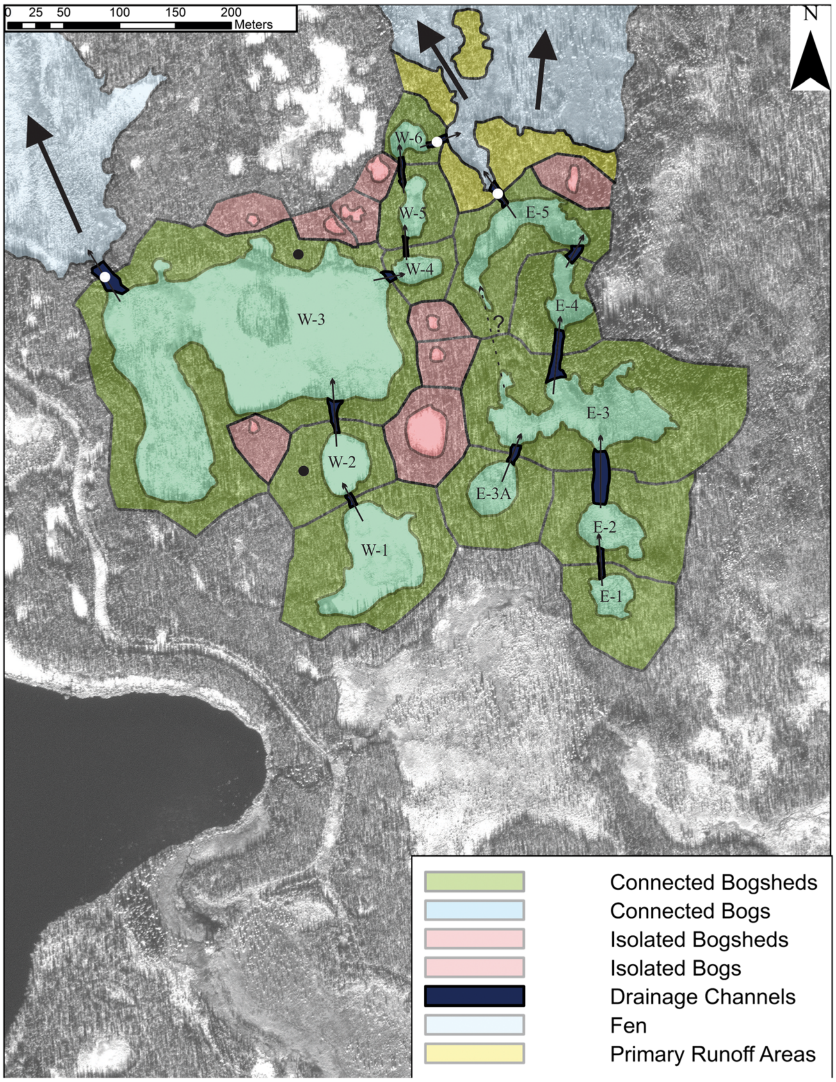

Figure 3. Map of two gauged bog series cascades at the Scotty Creek Research Basin. Note that bog W-3 has two outflows (W-3A to the northwest and another to bog W-4). Because of extremely flat topography in bogs, the drainage divide in bog W-3 is unknown. Black arrows indicate flow direction; white dots indicate location of weirs/flumes and black dots indicate location of tipping bucket rain gauges. Field observations indicate that there may be diffuse subsurface flow from the west edge of Bog E-3 into the southwest arm of Bog E-5 as indicated by the question mark

$2.0\,m^2$. Bog cascade flow direction is driven by the relative elevation of the bogs in the series such that those at higher elevations drain into those at lower elevations. Figure 4 depicts the side view of a bog cascade. Permafrost completely encircles each bog, with the exception of the thawed drainage channels which allow the bogs to transmit water. In some instances, the drainage channel has widened and merged two or more bogs, forming a complete connection (i.e. as in Figure 4C).

It is unknown how the drainage channels at the SCRB formed. We hypothesize that they are formed slowly, as the relatively warm bogs contribute energy that is supplied to thaw a channel through the intervening peat plateaus. Open water 'moats' exist along the perimeter of collapse scar features (Jorgenson *et al*., 2001) in discontinuous permafrost. Typically, the water table in bogs is below the ground surface (Hayashi *et al*., 2004) where the high porosity of the near surface *Sphagnum* mosses are able to provide thermal stability to the water in the bogs. Conversely, the moats that encircle the bogs are open water features. This increases the amount of net radiation at the surface of the moats (as opposed to the middle of the bog) and supplies energy to the system. These moats also receive additional energy inputs after rain events when the groundwater sitting atop the permafrost (i.e. supra-permafrost groundwater) is shed from the adjacent peat plateaus. A coupled transport of energy and mass occurs as water in the moat moves to the lowest elevation in the bog. We suggest that this influx of heat gradually erodes the permafrost and develops a

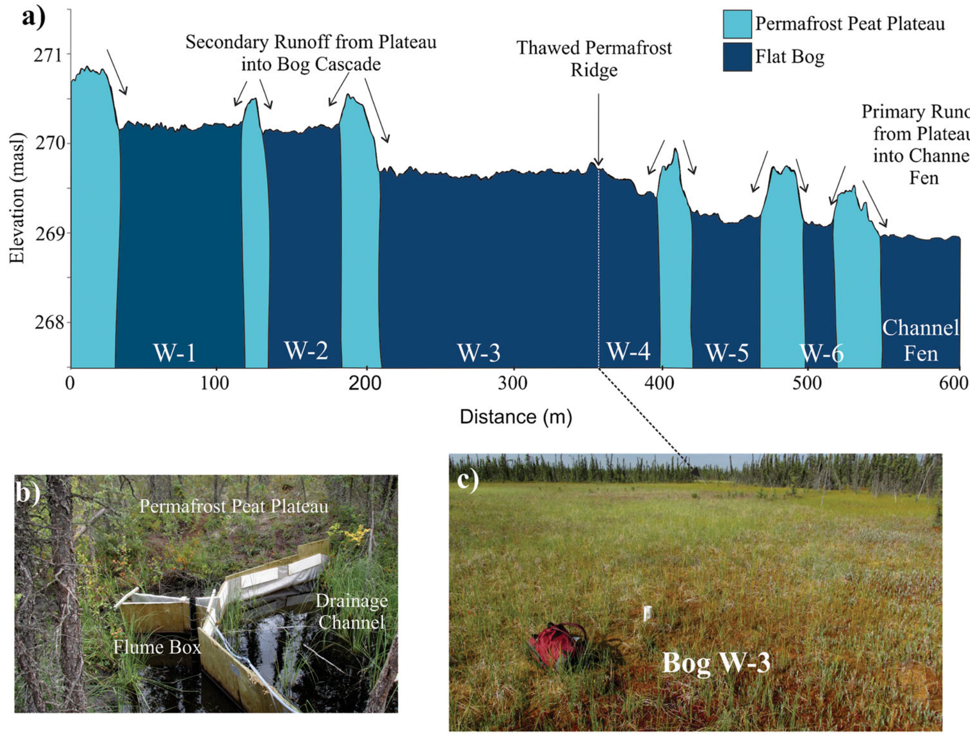

Figure 4. a) Cross-sectional area of bog cascade. Actual elevations were derived from a digital elevation model (DEM). The permafrost ridges encircle each bog with the exception of the drainage channels (not illustrated in diagram); b) photograph of a drainage channel and a flume box used for gauging discharge; c) photograph of bog W-3 in the foreground transitioning into bog W-4. There is no longer a drainage channel as the two bogs are now fully connected

drainage channel. Field observations have shown that the permafrost has completely degraded in the channels that connect flat bogs, as indicated by leaning trees on either side of the connections; however, the seasonally frozen layer has been observed to persist much later in the season (in some cases until early August) than the adjacent bogs which are typically thawed by mid-May. Very little is known about this mechanism behind permafrost thaw and it is not included in a review of 16 modes of permafrost thaw by Jorgenson and Osterkamp (2005).

METHODOLOGY

Two bog cascades (east and west) and their sub-catchments were initially identified from remotely sensed imagery (WorldView2) and a digital elevation model (DEM) derived from LiDAR imagery (Chasmer *et al.*, 2011). The catchments were then ground-truthed by inserting a steel rod into the ground until refusal along the perimeter of each bog cascade to ensure the presence of permafrost. Permafrost was found along the entire perimeter of the bogs within the cascades with the exception of the outflow drainage channels. Two terminal outflow channels were identified in the west cascade, and one terminal outflow channel was identified in the east cascade (Figure 3). The outflow channel W-3A is connected to a bog that was previously isolated from the main bog cascade network (as verified from a 1947 air photo; see Quinton *et al.*, 2011).

Sharp crested v-notch weirs were installed in the terminal connections in August 2012 to gauge outflow in 2013. In late August 2013 these were replaced with flume boxes, to measure larger volumes of water (Figure 4B). The base of the weirs and flumes were installed between 0.5 m and 1.0 m below the ground surface. It is assumed that during the spring freshet all water will flow through the weirs/flumes; however, it is acknowledged that subsurface flow beneath the weirs is possible as the thaw season progresses as the installation of the weirs represents a disturbance in the thermal regime of the system which may increase the vertical heat flux conducted beneath the weirs. Volumetric discharge measurements were obtained by capturing the flow through the weirs in a bucket or bag (dependent on volume) for a period of five and/or ten seconds (dependent on volume). Measurements were conducted a minimum of five times during each site visit to ensure accuracy. In three instances (i.e. peak flow in the east cascade in 2013) the bucket filled up in less than five seconds. In these instances the total time required to fill the bucket and the volume in the bucket were recorded to obtain discharge and at least ten measurements were taken. Measurements were obtained at least every other day at different stage levels between the onset of the freshet and when flow ceased in each channel. Manual stage measurements were taken during each measurement period with a metal ruler. Rating curves were then developed based on the stage–discharge relationship. The velocity of flow in the channels was too low ($<0.01\,\mathrm{m\,s^{-1}}$) to permit the use of the velocity–area method.

Runoff in 2013 was reported only using manual measurements because of low confidence in the computed hydrograph. Stage–discharge relationships are derived from power equations, indicating that small increases in stage can result in large increases in discharge. McLaughlin and Cohen (2011) note that if barometric pressure transducers are not installed in the same thermal environment as total pressure transducers (i.e. buffered by water temperatures) substantial differences in diurnal fluctuations may exist. As the barometric pressure transducer that we installed was subject to ambient temperature changes, our water level records displayed amplified diurnal signals (McLaughlin and Cohen, 2011), that when applied to our stage–discharge relationships produced high volumes of discharge that could not be verified by field measurements. As a result, 2013 discharge measurements are composed of only manual measurements; however these are still believed to be an accurate portrayal of the system.

In 2013, total pressure transducers with internal data loggers (Levelogger Gold F15/M5, Solinst Canada Ltd., Georgetown, ON, Canada, and HOBO U20 Water Level Data Logger, Onset Computer Corporation, Bourne, MA, USA) were installed in slotted stilling wells in the terminal connections and recorded at half hour intervals. A barometric pressure transducer (Solinst Barologger Gold) was also installed on-site. In the 2014 field season, vented pressure transducers (DCX-38 VG, Keller AG, Winterthur, Switzerland) were used instead of total pressure transducers and were directly connected to the flume boxes in the terminal connections. In both seasons, Solinst pressure transducers were installed in slotted stilling wells in bogs W-3 and E-3 to record water table level and recorded at half hour intervals. The water levels from the vented pressure transducers in 2014 displayed some anomalies, most notably large, sudden spikes between 03:00 and 07:00. Sudden spikes also appeared when the temperature approached and dropped below freezing. These spikes were likely caused by moisture condensing in the vent tubes and changing the pressure (McLaughlin and Cohen, 2011). There was no physically based reason for the water levels to spike during these periods, and as this was the time of day (i.e. 03:00 – 07:00) when air temperature was closest to the dew point these spikes were assumed to be a product of condensation and/or deposition. As a result, large spikes in water level originating during these time periods were deleted and linear interpolation was used to infill missing data. This linear interpolation may result in minor errors when estimating total discharge; however the error is estimated to be reasonably small (i.e. <5% of total discharge). Because of noise in the data, the east cascade hydrograph has been smoothed with a 3-h moving average. The west cascade data was less noisy and did not require smoothing.

Frost table and water table depths were monitored in all drainage channels in 2013 at least every other day in the spring (until 06 June) and at least every five days in the summer (07 June – 30 July). Frost table depth was measured daily at 0.5-m intervals in cross sections in each drainage channel by inserting a graduated steel rod into the ground until refusal. A level string was attached to two trees on either side of the channel and depths were measured down from this datum. Water table depth was measured in the same fashion but using a metal ruler instead. A level was used to ensure that the steel rod and ruler were level while measuring depth to the frost and water tables.

Snow surveys were conducted in late March, 2013 and 2014. Different land cover types (i.e. forests and clearings) accumulate different depths of snow water equivalent (SWE) over the course of a season. Snow surveys were completed for each bog in each cascade and on a minimum of three transects through the peat plateaus. Snow depth was measured using a metal ruler and SWE was measured using an Eastern Snow Conference snow sampler (ESC-30) in 2013 and a snow tube (GEO SCIENTIFIC Ltd., Vancouver, BC, Canada) in 2014. Each transect consisted of a minimum of five density measurements. Snow depth was recorded every 2, 5 or 10 paces depending on transect length. As snow density does not vary as much as snow depth, SWE was interpolated between points when density was not calculated. SWE was then calculated as an areally weighted mean over each sub-catchment.

Rainfall was measured using a tipping bucket rain gauge (0.2-m diameter, 0.35-m height) calibrated to 0.25-mm $\mathrm{tip^{-1}}$ connected to a data logger (CR 1000, Campbell Scientific Inc., Logan, UT, USA). Rain was measured at two locations on the study site (Figure 3). Precipitation data recorded at an Environment Canada monitoring station in Fort Simpson was used to compare the precipitation from the years of this study to 30-year climate normals (1981–2010; average annual precipitation: 387.6 mm).

Evapotranspiration (ET) from bog surfaces was estimated using the Priestley–Taylor method (Priestley and Taylor, 1972) at a bog within 1 km of the bog cascades:

$$ET = \alpha \frac{1}{\lambda} \left[\frac{s \cdot (Q^* - Q_g)}{s + y} \right] \quad (1)$$

where ET is total evapotranspiration (mm d^{-1}), α is the dimensionless Priestley–Taylor coefficient (see below), λ is the latent heat of vaporization (MJ kg^{-1}), s is the slope of the saturation vapour pressure – temperature curve (kPa °C^{-1}), y is the psychrometric constant (kPa °C^{-1}), Q^* (MJ m^{-2} d^{-1}) is total net radiation and Q_g (MJ m^{-2} d^{-1}) is the ground heat flux. Q^* was measured using a net radiometer (CNR1, Campbell Scientific), Q_g was measured using a ground heat flux plate (HFT3, Campbell Scientific), air temperature was measured using an HMP45C probe (Campbell Scientific) and all sensors were attached to a data logger (CR 10X, Campbell Scientific). The α coefficient was calibrated by comparison to actual ET calculated at an open-path eddy-covariance system installed on-site in May of 2014 (Helbig *et al*., in prep). The α value of 0.69 obtained in this study is within the range of 0.51 – 0.97 as reported by Gong *et al*. (2012), who compiled six different α values from studies in northern peatlands.

Storage capacity (m^3) is defined as the maximum amount of water a bog can hold before the outflow threshold is reached and is calculated as:

$$S_C = \theta \cdot D_M \cdot A_B \quad (2)$$

where θ is the porosity, D_M is the depth to underlying mineral soil (m) and A_B is the area of the bog (m^2). Porosity was estimated as the moisture content at saturation in a neighbouring bog and was set at 0.8. This value is consistent with other studies of northern wetlands (i.e. Spence *et al*., 2010). D_M was determined by use of a hand auger, where the depth was measured once mineral soil was found. Thaw of seasonal frost in bogs is heterogeneous and highly dependent on ground topography and moisture conditions. The frost table in bogs typically thaws within 1–3 weeks after snowmelt and has not been represented in calculations of storage deficit in this study. It should be noted that when frost is present, the storage deficit is bounded by the presence of seasonal frost and flow is restricted to the saturated layer above the frost table.

Soil moisture data was recorded at a neighbouring bog and plateau to compare the years of the study against moisture levels from previous years. Soil moisture values were recorded every minute and averaged every half hour using a soil moisture probe (bog: Hydra Probe II, Stevens Water, Portland, OR, USA; plateau: CS 615, Campbell Scientific) and connected to a datalogger (CR 10X, Campbell Scientific). This site was chosen because ongoing data collection has been occurring since 2008 and long term continuous soil moisture measurements are not available for the bog cascades. Both the bog and plateau soil moisture values were recorded at 10 cm below the ground surface.

RESULTS

Results from 2 years of monitoring indicate that secondary flow paths are most active during the spring freshet. Runoff from peat plateaus is rapidly shed to adjacent bogs at this time as storage is limited to the shallow thawed portion of the active layer. During this period, the surface of the bogs remains frozen and meltwater from the overlying snowpack begins to pond on the surface, typically serving to satisfy storage capacity in each bog. As the ice layer at the top of the bogs prevents water from entering the bog, infiltration is restricted (Granger *et al*., 1984). This is the period when water levels in the drainage channels begin to rise as water flows down gradient over the impermeable frozen surface of the bogs. This produces peak seasonal water levels in the drainage channels and induces surface flow as water moves through these features. During this period there is active flow through the drainage channels and the secondary contributing areas are fully connected. Once snowmelt has concluded, ET begins to contribute to the drawdown of the water table above the frozen bog surface. Between the months of May and August, ET occurs at an average rate of approximately 2.2 mm d^{-1}, with rates peaking around 4 mm d^{-1} near the summer solstice when available energy is at a maximum. Therefore, once the snow meltwater is rapidly flushed through the cascade system and into the fen, the degree of connectivity in the bog cascade decreases. As the thaw season progresses, ET rates higher than precipitation contribute to drying of the bogs so that progressively larger rainfall events are needed to satisfy the increasing storage deficit (Figure 5).

Channel development

Figure 6 shows the cross-sectional area of a drainage channel (connection W-6), along with the seasonal progression of the thawing of the frost table in 2013. The channel was not completely thawed until 26 July, which was 52 days after the adjacent (upstream) bog had thawed. The channels appear to preferentially thaw on the edges, leaving a residual frost bulb in the middle of the channel until mid-summer. All drainage channels were found to have taliks below the seasonal frost layer, which may facilitate subsurface flow; however the magnitude of the

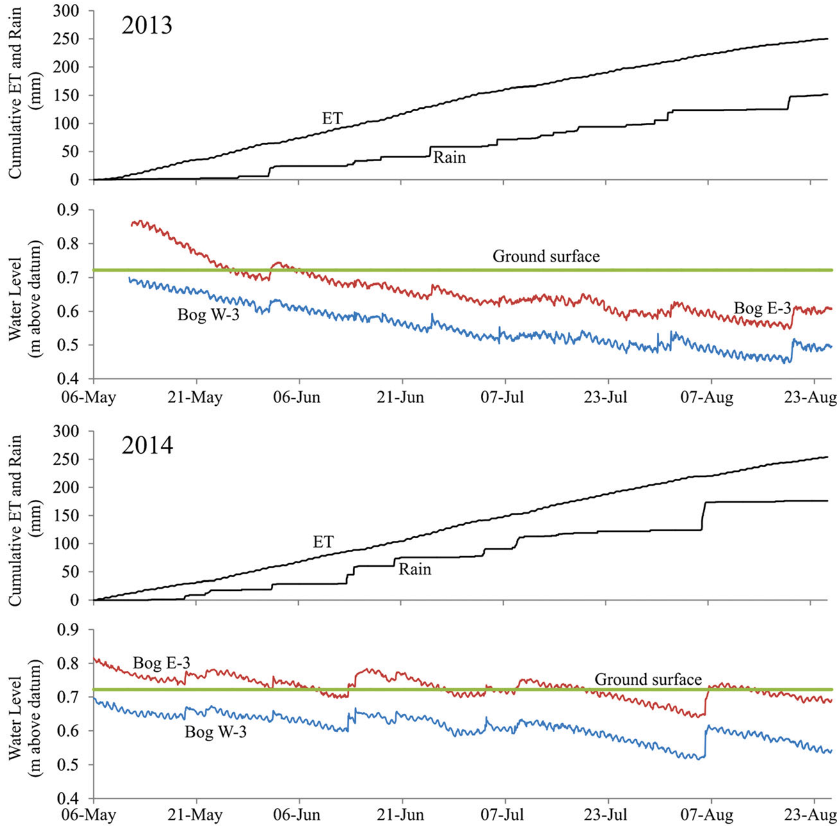

Figure 5. Cumulative evapotranspiration (ET) and rain as well as water table levels for bogs W-3 and E-3 in 2013 and 2014. The water table elevation is relative to an arbitrary datum and is shown in relation to the ground surface

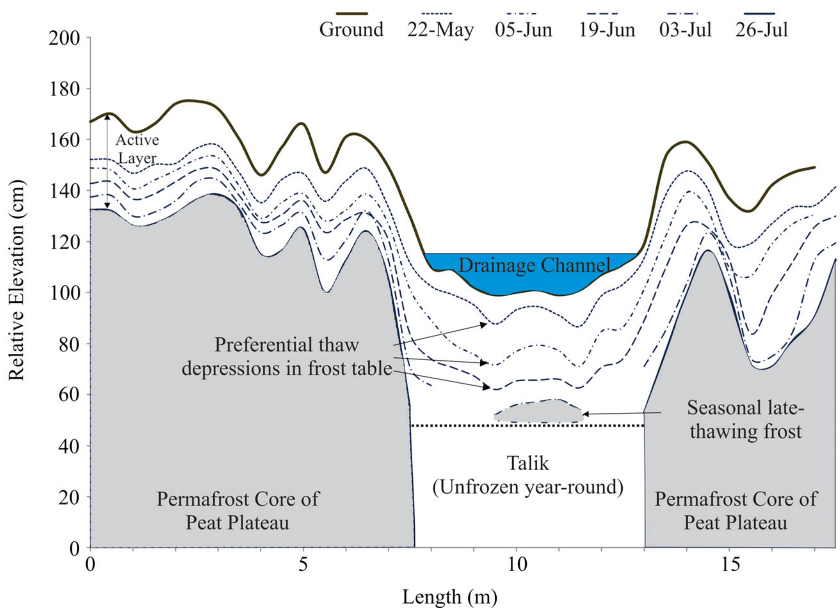

Figure 6. A cross section of a drainage channel (W-6). The dashed lines indicate the thawing of the frost table over the course of the spring/summer. Surface water exists primarily during the spring freshet when all bogs are connected. As the frost table lowers because of seasonal thaw, so too does the water table perched above it. Note that there is a 5× vertical exaggeration to denote the preferential thaw depressions in the frost table

flux is believed to be low. For example, Quinton *et al.* (2008) calculated that the hydraulic conductivity of peat at the SCRB decreases exponentially with depth. Below the surface and transition layers (uppermost 20 cm), there is a zone of uniformly low hydraulic conductivity (K_S = ~1 m d^{-1}).

Runoff produced from secondary contributing areas

Discharge from 2013 was estimated solely from manual measurements (Figure 7) because of low confidence in pressure transducer data (refer to methodology section). Although the diurnal cycles in the water level records in 2013 may be amplified (Figure 7B), the water level records are still reliable for determining the seasonal water level in the drainage channels. After flow in the drainage channels ceased (i.e. the water table dropped below the crest of the v-notch weir), field visits and water level records both indicate that surface flow did not resume over the course of the summer. The vented pressure transducers used in 2014 did not display the same diurnal anomalies that were observed in 2013, further verifying that diurnal fluctuations in 2013 were caused by imprecise barometric pressure readings.

Rating curves were developed in 2014 to create a continuous discharge hydrograph (Figure 8). Correlation between rating curves and manual measurements was strong (R^2 values between 0.92 and 0.96) for all three terminal drainage channels. Four summer rain events in 2014 were sufficient to re-introduce surface flow in the east cascade (represented by flow at connection E-5). The W-6 drainage channel exhibited a strong diurnal response to snowmelt with flows peaking at 19:00 on 28 and 29 April; however it was not able to sustain large flows after this period. The northwest outlet from bog W-3 (connection W-3A) produced steady discharge in response to snowmelt for about four days as water was slowly drained from the large bog. Peak snowmelt runoff in the east cascade (connection E-5) lagged the W-6 outlet by five days but maintained steady flows for two weeks after runoff commenced. The low flow period for three days before peak flow in the east cascade was verified by field measurements.

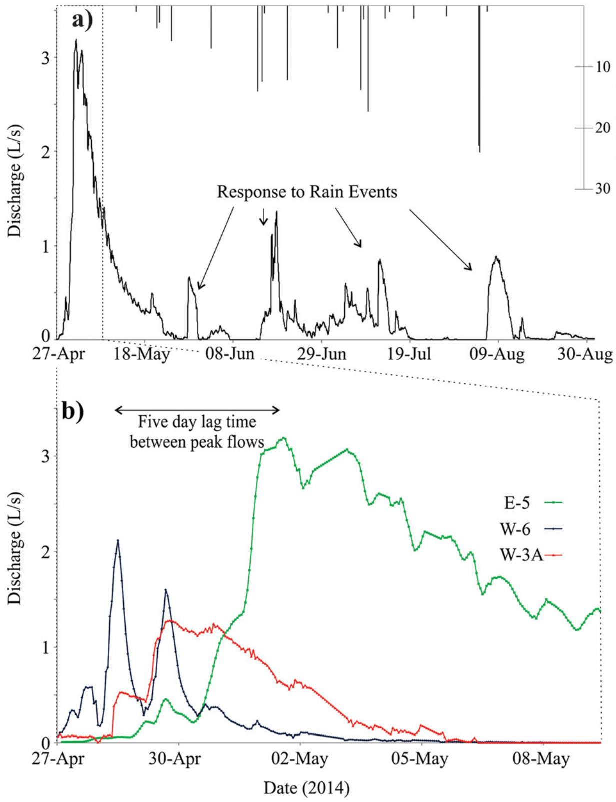

Figure 8. Total discharge in 2014. a) Hydrograph from the east cascade for the entire season. Snowmelt dominates the hydrograph, however response to rain events is evident; b) hydrographs for the three terminal connections during snowmelt, 2014 (no substantial rain events during this period). The peaks from channel W-6 occur around 19:00 each day and are likely a response to daily snowmelt

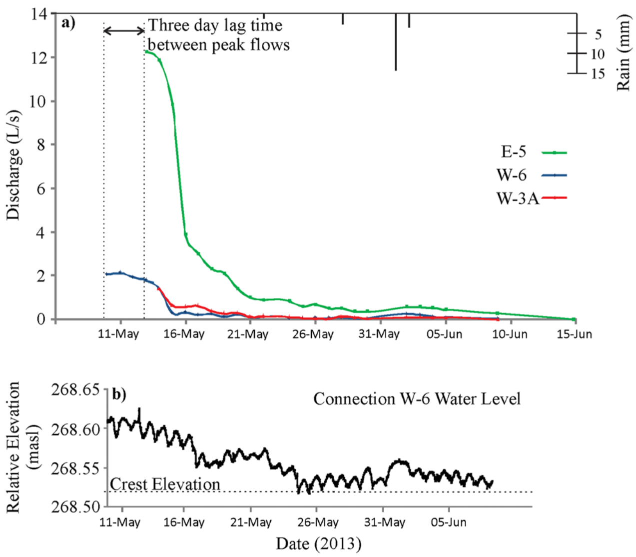

Figure 7. a) Discharge from the three terminal drainage channels in 2013. Channel E-5 was snow-choked until peak flows began May 13; b) water level records from channel W-6. The diurnal fluctuations are thought to be an artifact of the barometric pressure transducer being exposed to ambient air temperature fluctuations and not a response of the water level in the channel. The elevation of the crest of the v-notch is indicated by the dashed line

To quantify the error resulting from only using manual measurements in 2013, the same process was used to calculate discharge in 2014 and compared to the hydrographs derived from the water level recorders. The error between manual measurements and hydrographs produced from water level records in 2014 is 8%, which provides confidence in the 2013 manual measurements.

Impact on water budget

A notable result from this study is the difference in magnitude of runoff produced from two bog cascades that are on the same peat plateau complex. Table I shows the total discharge and runoff for the two bog cascades for 2 years. In 2013, the east cascade produced 74 mm of runoff, whereas the west cascade produced just 18 mm. In 2014, runoff decreased in both cascades because of drier

Table I. Total precipitation, discharge and runoff ratios for snowmelt and rain events in 2013 and 2014

Year	Cascade	Event/Date	Total discharge (m^3)	Total discharge (mm)	SWE (mm)	Rain (mm)	Runoff ratio
2013	West	Snowmelt	1621	18.1	172	24.3	0.09
		Season	1621	18.1	172	152.0	0.06
	East	Snowmelt	5025	73.8	170	24.3	0.38
		Season	5025	73.8	170	152.0	0.23
2014	West	Snowmelt	570	6.4	114	1.3	0.06
		Season	570	6.4	114	158.8	0.02
	East	Snowmelt	2049	30.1	115	1.3	0.26
		23-May-14	101	1.5	0	12.3	0.12
		13-Jun-14	187	2.7	0	23.0	0.12
		9-Jul-14	133	2.0	0	16.5	0.12
		6-Aug-14	309	4.5	0	46.5	0.10
		Season	3519	51.7	115	158.8	0.19

conditions. The east cascade produced 52 mm of runoff, while the west cascade produced 6 mm.

Total snow accumulation was very different between the 2 years of study. In 2013, there was an areally weighted mean of 172 mm of SWE, whereas 2014 received only 114 mm. Although 2013 had substantially more SWE than 2014, the snowmelt runoff ratios for both years were similar. The east cascade had a snowmelt runoff ratio of 0.38 in 2013 and 0.26 in 2014. This is comparable to the basin average at the SCRB (0.27; SD: 0.09) for the years 1996–2012 (data for the basin was either unavailable or incomplete for the years 2013 and 2014 at the time of publication). This indicates that the bogs of the east cascade are likely connected during the snowmelt period. The west cascade, producing much less runoff, had snowmelt runoff ratios of 0.09 in 2013 and 0.06 in 2014.

The period of peak flows immediately following snowmelt is assumed to be the period of full connectivity between bogs. After this, the bogs within the cascades become disconnected as surface flows among them cease. According to the Environment Canada meteorological station at Fort Simpson, both 2013 and 2014 received lower than average precipitation, while the years 2009 to 2011 were wetter than average. This allows for a comparison of soil moisture levels between wet and dry periods (Figure 9). The bog soil moisture data is particularly useful because it indicates when the soil was saturated or unsaturated at 10-cm depth. Quinton *et al.* (2008) show that the saturated hydraulic conductivity decreases by 2–3 orders of magnitude within the top 20 cm of the peat profile, and that the top 10 cm is a zone of uniformly high hydraulic conductivity (10 – 1000 m d^{-1}). Therefore, when the volumetric water content is at saturation at 10-cm depth, it can be inferred that runoff is more rapid. During the wet years, the soil at 10-cm depth was saturated for the majority of the summer and remained saturated until freeze-up. In 2012 soil moisture levels were at about 0.2 at freeze-up. 2013 was a year of high snow accumulation (172 mm SWE) and this snowmelt water was enough to maintain a water table in the top 10 cm of the bog peat profile until about 06 June. High volumes of runoff through the bog cascades during the 2013 freshet indicate that saturated soils at 10-cm depth may be a good indicator of runoff. This is further illustrated by the fact that there were no rain events large enough re-saturate the near-surface soil after 06 June and subsequently no runoff was observed from either bog cascade past this point. Freeze-up in 2013 also occurred when the water table was lower than the 10-cm threshold and was followed by a low snow year in 2014 (114 mm SWE). As a result, the water table was only within the top 10 cm of the peat profile for one day during the snowmelt season. This helps explain why the runoff ratios in 2014 were lower than in 2013. However, unlike 2013, subsequent rainfall events were sufficient to raise the water table back into the zone of high hydraulic conductivity later in the season.

The 2014 season contained four precipitation events of varying magnitudes (12, 23, 17 and 47 mm) which were able to reintroduce flow in the east cascade (Table I). The smaller events occurred in mid-summer (23 May, 13 June and 19 July), while the large event (47 mm) occurred in late summer (06 August). Figure 5 illustrates the effects of these rain events on water table level. The large rain event yielded comparable runoff (3.8 mm) as the smaller events (1.2 mm, 3.3 mm and 2.1 mm), likely because of further water table drawdown by this point in the season. The late season large magnitude event raised the water table in the bogs by about 100 mm but a portion of this was needed to satisfy storage deficits caused by cumulative evaporative loss over the course of the season. Figure 5 shows the importance of the water table elevation in relation to the ground surface. When overland flow occurs in the bogs (i.e. bog E-3), discharge was recorded at the cascade outlet. Overland flow did not occur in bog W-3 after the freshet in either year. Overland

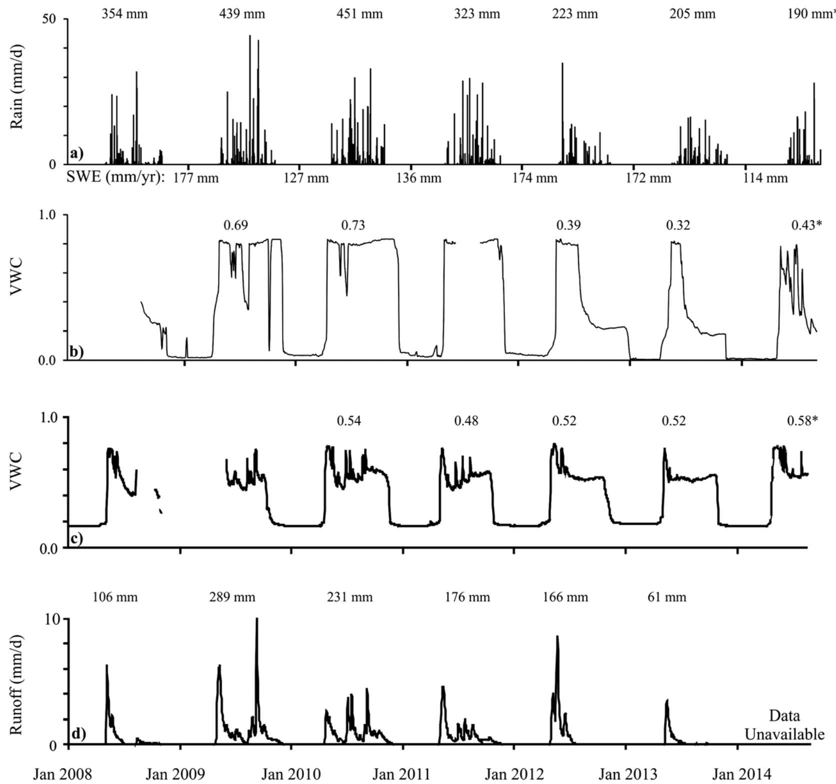

Figure 9. Precipitation (a), soil moisture (b, c) and runoff (d) data from 2008 to 2014 at Scotty Creek. Total rainfall, snow water equivalent (SWE) and runoff values are shown on each graph, while average soil moisture values for the thawed season are given. SWE values for each season are indicated below the x-axis as daily values of accumulation are not available. SWE values are listed as areally weighted averages for bogs and plateaus. Volumetric water content (VWC) values for soil at 10-cm depth in a bog (b) and an adjacent plateau (c) located about 800 m from the two bog cascades. Runoff was recorded by the Water Survey of Canada at the basin outlet. *indicates that data for the 2014 season is only available until the end of August

flow events in bog E-3 in 2014 align closely with discharge events at the outlet of the east cascade, showing that runoff is generated once storage capacity is satisfied.

DISCUSSION

Non-negligible flows from secondary systems

Previous studies (i.e. Quinton *et al.*, 2003) have suggested that bogs without an open connection to the channel fen are isolated and that they do not contribute to streamflow. Quinton *et al.* (2003) analysed runoff for four basins in the lower Liard River valley. They found a negative correlation between the percentage of bog areal coverage in a basin and basin runoff. In addition, they subdivided the SCRB basin into two sub-basins: the woodland dominated north sub-basin (2% bog coverage); and the wetland dominated south basin (15% bog coverage). They found that runoff ratios in the northern sub-basin were consistently higher, indicating that there were more storage deficits in the south sub-basin and by extension attributing this to the presence of more bogs. We have shown that drainage channels can hydrologically connect bogs to the channel fen and that these systems should not be ignored as runoff contributing areas. In some instances, particularly in response to melt events when the bog is frozen and infiltration is restricted, secondary runoff through bog cascades is comparable with the basin average. Therefore, bogs should not be universally classified as storage features, but rather as dynamic systems capable of contributing flow when moisture inputs are high. The cascades have also been shown to ‘re-connect’ in response to rain events and yield runoff ratios between 0.10 and 0.12 during these times, although it is noted that runoff ratios higher than this may be observed under wet conditions.

Quinton *et al.* (2009) conducted a detailed ground cover classification of 22 km^2 at the SCRB using IKONOS imagery. They found that the fractional total area covered by peat plateau–isolated bog complexes is 47%. The remaining 53% is comprised of open bogs, channel fens and lakes. Notwithstanding evaporative losses, and assuming there were no storage deficits, all precipitation falling on these landcover types can be conveyed to the basin outlet, as it will not be impounded by permafrost. Assuming that the peat plateau–bog complexes only contain isolated bogs and do not transmit water, the maximum water available for streamflow would be 53% of total precipitation (assuming spatially uniform distribution). If all isolated bogs have ephemeral connections and produce snowmelt runoff ratios equivalent to the east cascade (approximately 0.4), approximately an additional 20% of precipitation is made available to the channel fen during the spring freshet. This represents the maximum error that would be observed if estimates of streamflow are made without including secondary runoff contributing areas. Fully isolated bogs and corresponding bogsheds (which in the area of this study comprised 10% of the peat plateau–bog complex) will not have ephemeral drainage channels. In addition, not all bog cascades produce runoff ratios equivalent to the east cascade, therefore this is likely an upper limit of the estimated impact. The actual quantity of additional precipitation from secondary runoff contributing areas made available to the channel fen network during snowmelt is likely between 5 and 15%, but may increase up to 20% as the presence of isolated bogs is reduced with changing climate. This analysis is intended to provide an estimate of the amount of runoff that would be neglected if secondary runoff producing areas are not accounted for in numerical models. Further analysis of the per cent cover of peat plateau and isolated bog complexes is needed over an array of sub-watersheds to more accurately quantify the influence of secondary runoff pathways on a basin scale.

Controls on runoff generation

Runoff generation theories designed for temperate environments are not necessarily transferrable to permafrost landscapes. For example, field observations have shown that some assumptions for the ‘variable source area’ concept (Hewlett and Hibbert, 1967) cannot be satisfied on permafrost slopes (Woo, 2012 page 297). Spence and Woo (2006) present an ‘element threshold’ concept to better describe runoff processes in permafrost terrain. This process of runoff generation is a conceptual upscaling of a ‘fill-and-spill’ concept presented by Spence and Woo (2003) in a soil-filled valley. The concept was developed from studies in a basin in the Canadian Shield, where runoff is often dependent on basin physiography of the underlying bedrock. According to the element threshold concept, runoff from individual elements is released only when storage capacity in that element has first been exceeded. We suggest that the element threshold concept can be applied to the movement of water through bog cascades in discontinuous permafrost. The storage capacity of each bog must first be satisfied before high volume surface and near-surface flows can be transmitted to the next bog. If the water level in one bog is below the outflow threshold, it intercepts all flow generated from each upstream bog. Storage capacity is directly proportional to the surface area of a bog, indicating that larger bogs require greater inflow volumes to exceed their storage capacity (shown in Table II).

A marked difference in runoff between two cascades that are adjacent to each other and on the same peat plateau complex indicates that there is substantial spatial heterogeneity in bog cascades. Stark differences in runoff ratios are found because of dynamic and heterogeneous contributing areas. Large bogs with corresponding large storage capacities and small contributing areas (Table II) may be more effective in retaining their water than smaller bogs with large contributing areas. For example, bog W-3 is the middle bog in its cascade and has a plateau-to-bog ratio of 0.64. Conversely, the lowermost three bogs in that same cascade have a plateau-to-bog ratio of 2.49. The east cascade has a similar plateau to bog ratio of 2.95. If the effective contributing area for the outlet of bog W-6 is contracted to include just the lower three bogs in the cascade (implying that the large bog W-32 does not transmit water), the snowmelt runoff ratios are 0.58 in 2013 and 0.20 in 2014, which is close to the snowmelt runoff ratio of the east cascade. The higher runoff ratio of 0.58 in 2013 indicates that bog W-3 may have been able to transmit subsurface water during the freshet of that year. The runoff ratio of 0.20 in 2014 is similar to the runoff ratio in the east cascade for 2014 (0.26), indicating that bog W-3 may have been disconnected from the bog cascade during the snowmelt runoff period of 2014. As soon as bog W-3 is no longer able to transmit water, the entire contributing area upstream of that bog also no longer contributes. It should also be noted that there is a substantial lag time (difference between peak rainfall/snowmelt and peak flow) for runoff in the east cascade compared to the west cascade. During snowmelt this lag time has been observed to be up to five days. This lag time is likely caused by a combination of two factors: (1) snow-choked channels preventing the transmission of water during the spring freshet; and (2) the requirement for each bog to exceed storage capacity before water can be transmitted downstream. Snow-choked channels have been shown to delay outflows from lakes in subarctic environments by up to 10 days (FitzGibbon and Dunne, 1981) and also to raise the threshold required to generate outflow (Woo and Mielko, 2007).

Both years in this study period received lower than average precipitation, while a very dry summer in 2013 was compounded by very little snowfall in 2014. As a result, the spring freshet in 2014 produced the lowest stage at the Water Survey of Canada gauge since level monitoring began in 2002. The two subsequent dry years allowed for the depression storage capacity in bog W-3 to not be satisfied for a period long enough to contribute sustained flows to the cascade outlet in 2014. It can be

Table II. Geometric properties and storage capacities for each bog in the two bog cascades

	Bog W-1	Bog W-2	Bog W-3	Bog W-4	Bog W-5	Bog W-6	West Cascade	Bog E-1	Bog E-2	Bog E-3	Bog E-3A	Bog E-4	Bog E-5	East Cascade
Bog area (m^2)	5819	1966	34 421	953	1069	829	45 056	1240	2058	7099	1730	1616	3503	17 246
Plateau area (m^2)	9862	5438	22 092	1930	3259	1901	44 483	5818	8687	18 509	5798	4481	7586	50 879
Bogshed area (m^2)	15 681	7404	56 513	2883	4328	2730	89 539	7058	10 745	25 608	7528	6097	11 089	68 125
Plateau:bog ratio	1.69	2.77	0.64	2.03	3.05	2.29	0.99	4.69	4.22	2.61	3.35	2.77	2.17	2.95
Peat thickness (m)	3.9	3.7	3.7	2.8	2.2	2.85		1.95	1.7	1.5	2.05	1.3	2.1	
Storage capacity (m^3)	18 155	5819	101 886	2134	1881	1891	131 766	1934	2799	8519	2837	1681	5885	23 655

surmised that large bogs with small contributing areas require substantial inputs to satisfy their storage capacity. This is shown by water table levels never exceeding the ground surface in bog W-3, whereas bog E-3, with a larger proportional contributing area, had a water table higher than the ground surface following three rain events in 2014 (Figure 5). The findings here are consistent with those of Mielko and Woo (2006) who studied runoff generation in a headwater lake in the Canadian Shield, where lake outflow can be considered analogous to bog outflow. These authors found that antecedent lake storage conditions and ratio of catchment to lake area are two important factors to generate outflow. It should be noted that when this study was scaled up to the catchment scale, a large basin to lake ratio was not a critical consideration in lake outflow (Woo and Mielko, 2007). In the same region, Phillips *et al.* (2011) found that some landscape elements (i.e. lakes or bogs) can have a 'gatekeeper' effect on hydrologic connectivity. In this study it appears that bog W-3 acts as a gatekeeper and is able to disrupt upstream connectivity when it is inactive. Under wet conditions it is hypothesized that bog W-3 will remain active longer into the summer and serve to attenuate runoff in a manner similar to active gatekeepers as described by Phillips *et al.* (2011).

Challenges in predicting secondary runoff

The hydrometric network in northern Canada does not have sufficient capacity to understand the implications of climate change on northern water resources (Spence *et al.*, 2007). As a result, there has been a concerted effort to increase our ability to monitor northern sites remotely (i.e. Töyrä and Pietroniro, 2005). For example, the launch of the twin Gravity Recovery and Climate Experiment (GRACE) satellites allows for changes in water storage to be detected from space (Tapley *et al.*, 2004); a very powerful tool to study the implications of permafrost thaw on the basin water balance. Unfortunately, connected bog regimes cannot be accurately identified using remote sensing devices. For example, Quinton *et al.* (2003) used IKONOS (resolution 4 m × 4 m) and Landsat (resolution: 30 m × 30 m) imagery to classify different land-cover types at the SCRB. In their analysis, they determined that if a bog had a pixel in direct contact with a channel fen, this would be classified as a connected bog. As a result, all bogs in both cascades discussed in this study were incorrectly determined to be isolated and therefore were thought to be storage features. LiDAR based DEMs can be used to delineate the presence of wetlands in the SCRB, as their ground elevation is lower than that of the surrounding peat plateaus. Although the DEM is useful for identifying some drainage channels, DEM analysis inaccurately produces others that are not verifiable by ground-truthing. Therefore, a process-based understanding of these systems is necessary to drive conceptual and numerical models.

Considering secondary contributing areas at the basin scale becomes further complicated when considering the density of old roads and seismic lines that spread across the landscape. These are linear features where the forest canopy has been removed, resulting in permafrost thaw and ground subsidence (Williams *et al*., 2013). These cut lines act as drainage features and may intersect with bog cascades. Williams *et al*. (2013) calculated the flow out of a drainage channel connecting a bog and a seismic line and found peak flow rates of about $5\,\mathrm{L\,s^{-1}}$. The hydrological implications of linear cut lines are the subject of other on-going studies; however it is worth noting that their presence makes understanding the hydrology of bog cascades more difficult as catchments are not easily identifiable.

Permafrost thaw in the SCRB is rapidly changing the landscape (Quinton *et al*., 2011) and in doing so is changing the basin hydrology. We hypothesize that there is a relationship between the plateau-to-bog ratio and the runoff ratio for a bog cascade; however our limited sample size of two cascades limits the confidence in this relationship. If more data is collected and a relationship can be established, this may be an effective tool in predicting the impacts of land cover change resulting from permafrost thaw. As permafrost thaw continues to manifest itself across the landscape, both the density and size of secondary contributing areas is expected to increase. Basin contributing areas are expected to grow as there are less isolated bogs across the basin. Developing improved conceptual and numerical models of secondary contributing areas is crucial to understanding northern water resources.

CONCLUSIONS

Here, we present evidence that bogs are capable of acting as flow-through features in a thawing discontinuous permafrost environment. Drainage channels that connect bogs are capable of forming a cascade of bogs that can extend over a large contributing area, effectively expanding the runoff contributing area in a basin. These cascades tend to be fully active during snowmelt and conditionally active to a lessening degree over the course of summer. Heterogeneities in the landscape result in spatially disparate runoff regimes within only tens of metres. Over the two year study period, one bog cascade produced 125 mm of runoff while the adjacent cascade produced just 25 mm. We propose that an 'element threshold concept' runoff regime governs discharge, and that large bogs with high storage capacities and small contributing areas may serve to limit outflows, especially during dry periods. Peak flows occur during the spring freshet, when all bogs in the cascade contribute to flow. As runoff and evaporation contribute to water table drawdown, the bogs become disconnected from each other and surface flow ceases. Surface flow through the drainage channels has been observed to re-start in response to summer rain events, producing runoff ratios between 0.10 and 0.12. Secondary contributing areas produce flows sufficient enough that warrant their processes to be incorporated in conceptual and numerical models. For example, ignoring these flows during melt seasons would lead to underestimates of flows between 5 and 15% with the possibility to increase up to 20%. As permafrost continues to thaw, these contributing areas are expected to become more prominent across the landscape. Identifying secondary contributing areas using remotely sensed imagery is not accurate and therefore field-intensive programmes are required to provide a better understanding of these systems.

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