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Permafrost thaw induced changes to runoff generation and hydrologic connectivity in low-relief, discontinuous permafrost terrains

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by

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DISSERTATION

Submitted to the Department of Geography and Environmental Studies

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Declaration of Co-Authorship/Previous Publications:

Co-Authorship Declaration:

I declare that I am the principal author of this thesis and all research within is the product of my own work. Chapters 2, 3, and 4 represent the main body of the thesis and contain material that has been published (Chapters 2 and 3) and material that is set for publication (Chapter 4). This dissertation is the result of collaborative work with the co-authors listed on each publication. All co-authors provided expertise and edits for the chapters in which they were involved. Dr. William Quinton is a co-author on all three chapters and has provided invaluable support in conducting field measurements (2013-2017) and through discussion of hydrological processes. Dr. James Craig is a co-author on Chapters 2 and 3 and provided expertise in developing conceptual frameworks and discussion of incorporating field data into numerical models. Dr. Masaki Hayashi is a co-author on Chapters 2 and 4. Masaki provided strong expertise on field processes and installed many of the sensors used in the analysis in Chapter 4. Dr. Oliver Sonnentag is a co-author on Chapter 3. Oliver was involved in field work and infrastructure development, and provided valuable feedback on processes. Jessica Hanisch is a co-author on Chapter 3 and Élise Devoie is a co-author on Chapter 4. Jessica and Élise aided in the collection of field data and contributed to the development of the ideas presented in the respective chapters.

Previous Publications:

Chapters 2 and 3 of this thesis have been previously published in *Hydrological Processes*. Chapter 4 is set for submission to the *Journal of Geophysical Research – Earth Systems*. Appendix A consists of a reprint of a manuscript submitted to the 2016 CGU Annual General Meeting. This manuscript is not included in the thesis as some ideas presented are similar to those presented in Chapter 4, however it also includes additional information and data about the routing of subsurface water and serves as a good reference point for readers.

ABSTRACT:

Recent climate warming in northwestern Canada is occurring at an unprecedented rate in recorded history and has resulted in the widespread thaw of permafrost. Where present, permafrost exerts a significant control on local hydrology, and disappearance of permafrost threatens to change the hydrology of northern basins. In the peatlands that characterise the southern distribution of permafrost in low relief terrain, permafrost takes the form of forested peat plateaus and is interspersed by permafrost-free wetlands (*i.e.* channel fens and flat bogs). Previous field studies have found that channel fens serve as the drainage network and route water to the basin outlet, whereas flat bogs have been viewed primarily as storage features. Wetland expansion in response to permafrost thaw can transform the primary hydrologic function of flat bogs from storage units to runoff-producing units by removing the relatively impermeable permafrost barrier that encompasses them. As a result, permafrost thaw has the potential to greatly increase the runoff contributing area when large storage features form hydrological connections with the basin drainage network. It has been well documented that permafrost thaw in this region results in the loss of forest and a concomitant expansion of wetlands, however the hydrologic response of these changes is poorly understood.

Stream flow records in four Water Survey of Canada gauged basins (152 - 2050 km^2) in the lower Liard River valley were analyzed to determine the impact of permafrost thaw-induced land cover change on basin runoff. Annual runoff between 1996 and 2012 increased by between 112 mm and 160 mm and these changes were significant in all four basins ($p>0.05$). Changes to stream flow were assessed using the Mann-Kendall non-parametric test and the Kendall-Theil robust line. Permafrost thaw between 1977 and 2010 was quantified by comparison of historical aerial photographs and high-resolution satellite imagery (World View 2) over a 6 km^2 area of

interest, where changes in tree-covered terrain were used as a proxy for permafrost loss. It was found that land cover change from forest to wetland was the most important factor contributing to the increases in runoff (37 -61 mm), and that basins with a relatively high cover of flat bogs were subject to the largest increases in runoff.

This analysis examined increases in runoff contributing area when a direct connection was formed between wetlands. Field studies have indicated the presence of ephemeral drainage channels connecting flat bogs within a peat plateau-bog complex. These drainage channels cut through the peat plateaus and create a series of cascading bogs that ultimately discharges into the channel fen. The bog cascades can greatly increase the runoff contributing area of a basin when the cascade is hydrologically active. To investigate the transport of water through these features, two bog cascades were instrumented with sharp crested v-notch weirs and cut-throat flume boxes in 2013 and 2014. Within the peat plateau-bog complex, the two cascades had markedly different plateau:bog ratios, and therefore different contributing areas. Runoff between the two cascades varied significantly with one cascade producing 125 mm of runoff over the two year period and the other producing only 25 mm. Both cascades were active during the snowmelt period of each year, however only the cascade with the higher plateau:bog ratio produced runoff in response to rain events. It is proposed that the 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. This work suggests that neglecting to represent these cascades of connected bogs in numerical models can underestimate basin stream flow by between 10 and 15%.

At the southern distribution of discontinuous permafrost, the rate of permafrost thaw has increased in recent years. This research demonstrates a mechanism whereby permafrost thaw

may show a non-linear response to warming air temperatures. Measurements of active layer thickness (ALT) are typically taken at the end of summer and inherently assumed to be analogous to maximum thaw depth. By definition, the active layer is the layer above permafrost that thaws in the summer and freezes again in winter. In Subarctic Canada, field measurements conducted at the end of winter found that the entire thickness of ground atop permafrost does not entirely re-freeze. This results in the formation of a thin talik between the frozen active layer and permafrost, and indicates that ALT must be measured by the depth of re-freeze. As talik thickness increases at the expense of the underlying permafrost, ALT is shown to simultaneously decrease. This suggests that the active layer has a maximum thickness that is controlled by the amount of energy lost from the ground to the atmosphere during winter. Vertical permafrost thaw was found to be significantly greater in areas with taliks (0.07 m year^{-1}) than without (0.01 m year^{-1}). Furthermore, the spatial distribution of areas with taliks increased between 2011 and 2015 from 20% to 48%, a phenomenon likely caused by an anomalously large ground heat flux input in 2012. Wide-spread talik development can therefore expedite permafrost thaw and add further complexity to the observed changes to basin hydrology.

Keywords: permafrost thaw, runoff contributing area, hydrologic connectivity, wetlands, bog cascade, talik, ground heat flux, organic soils.

ACKNOWLEDGEMENTS:

First and foremost, I would like to thank my supervisor, Dr. Bill Quinton. The support that Bill has provided over the course of my PhD is more than I could have ever anticipated or imagined. Bill's enthusiasm about hydrology and field work has been very contagious. He has shown me what it takes to be a scientist, how to ask questions, and what is involved in a designing and implementing a field research programme. I have enjoyed our many conversations that have often lasted late into the evenings, whether they were about science or hockey. I would also like to thank Bill's family: his wife, Helen, and his three daughters, Julia, Laurie, and Mary, who have all made it possible for Bill to be as invested in his research as he is. They have treated me like family over the past five years and I am extremely grateful for their support.

This research would not have been possible without generous funding support from the W. Garfield Weston Foundation, the Natural Sciences and Engineering Research Council of Canada (NSERC), the Ontario Graduate Scholarship (OGS), the Northern Scientific Training Program (NSTP) and Wilfrid Laurier University (WLU). The WLU – Government of the Northwest Territories (GNWT) Partnership and the Changing Cold Regions Network (CCRN) have also provided support for this research. Further support was provided through NSERC Discovery grants awarded to Bill Quinton.

I would like to thank the members of committee, Drs. James Craig, Masaki Hayashi, Oliver Sonnentag and Phil Marsh for working with and teaching me over the past five years. I have cherished the opportunity to learn from you and to navigate the scientific process. I would also like to thank Drs. Aaron Berg and Jennifer Baltzer for working with me, and sharing insights at Scotty Creek over the course of my studies.

The research presented here is the result of five extended field seasons in the NWT. The work was made possible by the support of colleagues in the field who were able to make the sunny days some of the best of my life and helped to push through the rainy ones. Unfortunately there are too many people to list here, but to everyone who I have met and worked with at Scotty Creek -- I will cherish those memories forever. I owe a special thanks to everyone who worked with me as collaborators and field assistants, including John Gordon, Stacey Van Opstal, John Coughlin, Emily Haughton, Lindsay Freeman, Lindsay Stone, Jessica Hanisch, Manuel Helbig, Nick Lillie, Karoline Wischnewski, Allison McManus, Rebecca Warren, Jared Simpson, Elyse Mathieu, Caren and Clara Ackley, Michael Braverman and Élise Devoie. Whether it involved sacrificing dry feet for a measurement, or braving days with clouds of black flies and mosquitoes, I am thankful for all the time and effort that you have dedicated to the project.

Conducting field work in remote locations is very challenging logistically. The Scotty Creek project would not be possible without the support of many people in the Northwest Territories. In Yellowknife, Pete Cott and Pierre Berube have provided support in preparing and fixing field equipment, trucks, trailers and snowmobiles. Wayne and Lynn McKay at Checkpoint have been incredibly generous in offering accommodations, storage and field support. They have maintained snowmobile trails and infrastructure in camp during the winter months. Their contribution to the Scotty Creek camp is invaluable.

I would also like to thank my family and friends who have stood by my side for the last five years. Your endless support and encouragement have allowed me to get to where I am today. I have missed many birthdays and holidays because of time spent in the field. I appreciate the time that you have spent listening to talks and proof-reading my work. My parents have been wonderful and incredibly supportive over the past five years, and I owe an immense amount of

gratitude to them. Last but not least, to my girlfriend Cassandra, thank you for sticking by my side as I disappear for three months at a time. I know that it can be difficult, but knowing that you support me means more than you can imagine.

STATEMENT OF ORIGINALITY:

I hereby declare that the work presented herein is the result of my own investigations. This work has not been submitted in whole, or in part, for recognition at Wilfrid Laurier University or any other University. All methodologies and ideas presented in this paper are either original, or have been properly cited and acknowledged.

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CHAPTER 1:
General Introduction

1.1 INTRODUCTION AND BACKGROUND:

Permafrost, ground that remains at or below 0°C for two or more consecutive years, underlies over half of the Canadian land surface and almost 25% of the exposed terrestrial land surface in North America (Bonan, 2008). In cold climates, where mean annual air temperature (MAAT) is < -8.5°C, permafrost coverage is continuous and typically has a thickness of 100 to 1000 m (Prowse, 1990). As MAAT increases, permafrost bodies become thinner and discontinuous in nature, although it should be noted that permafrost thickness is also highly dependent on the geothermal gradient, presence/chemical composition of sub-permafrost groundwater and proximity to surface water bodies (Kudryavtsev, 1959). In the zone of discontinuous permafrost, permafrost underlies between 10% and 90% of the landscape and can be further classified as widespread or sporadic discontinuous permafrost (Prowse, 1990). At the southern edge of the permafrost distribution, permafrost manifests in isolated patches below the landscape, where site-specific conditions allow for its preservation. MAAT offers an approximation for permafrost distribution (Chadburn *et al*., 2017; Åkerman and Johansson, 2008), but is not the only determining factor for its presence. Other factors influencing permafrost distribution include topography (Woo, 2008), aspect (Carey and Woo, 1999), snow cover (Kokelj *et al*., 2017; Johansson et al., 2013), soil type, vegetation (Jean and Payette, 2014; Jorgenson *et al*., 2010), and local hydrological factors (Sjöberg *et al*., 2016).

The active layer is the layer of ground above permafrost that thaws in the summer and freezes again in the winter (ACGR, 1988; Muller, 1947). In some cases, especially near the southern edge of discontinuous permafrost, a talik (layer of unfrozen ground) exists between the active layer and permafrost. Muller (1947) defines the combined thickness of the active layer and talik as the suprapermafrost layer. As the majority of hydrological processes in permafrost terrains are constrained to the active layer (and talik, if present), it is important to document changes to the thickness of this layer over time (Tarnocai *et al*., 2004). Large scale, collaborative efforts, such as the Circumpolar Active Layer Monitoring Programme (CALM) have developed to measure and document changes to active layer thickness (ALT) over the circumpolar north (Brown *et al*., 2008).

Decreases in temperature below 0°C dramatically reduce the water potential and the hydraulic conductivity (Hayashi, 2013; Burt and Williams, 1976) of soil, demonstrating the fundamental hydrologic controls that permafrost exerts on soils. When soil water freezes, the diameter of air space within pores decreases, which in turn results in a decrease in soil permeability to the second power (Freeze and Cherry, 1979). As soil temperatures drop further below 0°C and remaining liquid pore water turns to ice, transmission of water through permafrost bodies declines to a point where they become relatively impermeable. If, however, variable amounts of water remain in the liquid form at temperatures below 0°C (*e.g.* in systems with high salinity), groundwater flow can still be an effective transport mechanism in permafrost soils (see review by Kurylyk *et al*., 2014). The hydrology in permafrost terrains is intrinsically linked to surface and sub-surface energetics, as storage and volume of runoff is dependent on seasonal thaw depth (Quinton and Carey, 2008; Young and Woo, 2005). If permafrost thaws and changes the spatial distribution or thickness of the permafrost, substantial changes to the hydrological regime can be expected (Quinton and Baltzer, 2013; St. Jacques and Sauchyn, 2009; Marsh and Neumann, 2003). Process-based investigations are required to understand how these changes may manifest themselves at the landscape scale.

Climate warming in northern Canada has occurred rapidly during the past 100 years with the largest temperature increases occurring during the winter months (Vincent *et al*., 2015; McBean *et al*., 2005). Increasing temperatures threaten to degrade both the areal extent and thickness of permafrost. Modelled and measured rates of permafrost thaw suggest that the global areal extent of permafrost decreases by between 1.8 and 4.0 million km^2 per 1 °C of increase in mean annual global average temperature (Chadburn *et al*., 2017; Koven *et al*, 2013). In North America, rapid permafrost thaw has been documented in areas of sporadic and discontinuous permafrost (Jepsen *et al*., 2013; Quinton *et al*., 2011; Mazhitova *et al*., 2008; Kwong and Gan, 1994). Thawing of permafrost, especially in areas with high ice content, can fundamentally change the physical structure of the ground, and in turn, affect the hydrology (Sjöberg *et al*., 2013; St. Jacques and Sauchyn, 2009; Woo, 1986), ecology (Sniderhan and Baltzer, 2016; Kemper and Macdonald, 2009), and gas exchange (Helbig *et al*, 2016a; Johnston *et al*., 2014; Turetsky *et al*., 2007).

Permafrost thaw is likely to cause substantial changes to surface and subsurface hydrology, especially in areas of discontinuous or sporadic permafrost. Where permafrost is discontinuous, baseflow is dependent on the proportion of the basin underlain by permafrost (Jones and Rinehart, 2010; Rouse *et al*., 1997). Some studies have hypothesized that the thawing of permafrost can re-activate deep groundwater pathways and increase baseflow contributions (St. Jacques and Sauchyn, 2009; Walvoord and Striegl, 2007). In areas with high relief, aspect plays a determining role in the presence of permafrost, and in turn, on local hydrology. Carey and Woo (1999) found that north-facing slopes underlain by permafrost produce significantly more runoff than south-facing slopes, where almost all precipitation inputs are either lost through evapotranspiration or retained as storage. Many other studies comparing hillslopes with permafrost to those without found similar results (Bolton *et al*, 2000; Chacho and Bredthauer, 1983; Kane *et al*., 1981), underscoring the controls that permafrost exerts on local hydrology. In areas of discontinuous or sporadic permafrost with low relief, permafrost is not a function of aspect, but of the insulating properties of near surface soils and/or vegetation (Brown, 1970).

The scope of this thesis is limited to low-relief sporadic and discontinuous permafrost environments where permafrost is preserved by the highly insulative properties of peat. Water movement and storage in flat, low-lying areas of discontinuous permafrost is poorly understood. Sparse hydrometric and meteorological monitoring networks in Canada's north hinder our ability to understand hydrological processes (Spence *et al*., 2007). Near the southern edge of the permafrost distribution, permafrost exists almost exclusively in peatlands, taking the form of permafrost cored palsas or peat plateaus (Smith and Riseborough, 2002; Vitt *et al*., 1994). The peatlands in northwestern Canada (around 60°N) were first established as fens about 8000 to 9000 years before present (BP), after the retreat of the Laurentide ice sheet (Zoltai, 1995, 1993). Bog development was initiated about 5000 years BP. The first evidence of permafrost in this region appears to have occurred about 3700 years BP (Zoltai, 1993). Since then, there has been a natural pattern of permafrost aggradation and degradation in northern peatlands (Zuidhoff and Kolstrup, 2005; Zoltai, 1993; Seppälä, 1988; Payette *et al*., 1975). Permafrost thaw below peat plateaus can occur on the periphery of plateaus, where adjacent wetlands expand outward, or on the interior of the plateau through the formation of collapse scar bogs and fens (Robinson and Moore, 2000; Zoltai, 1993). Collapse-scar features form when anthropogenic or natural factors alter the ground surface energy balance and increase conductive heat transport to the underlying permafrost.

Field observations at Scotty Creek, NWT indicate that in the early stages of permafrost thaw, a topographic depression forms on the permafrost table, where subsurface water accumulates, as directed by the local hydraulic gradient. Due to the high porosity of peat, the bulk thermal conductivity is highly sensitive to moisture content increases which initiate a positive feedback, allowing for a higher conductive energy flux from the ground surface to the thawing permafrost table, and accelerating the thaw rate. Ground subsidence also occurs in response to permafrost thaw as a result of volumetric contraction of the substrate when pore ice melts, leading to localized flooding of the ground surface. Eventually, *Spaghnum* mosses become the dominant vegetation type and a collapse-scar bog is formed. Peat accumulation leads to vegetation succession allowing *S. angustifolium* and *S. fuscum* grow above the water table, forming hummocks that can be up to 40 cm higher than the water table (Bauer and Vitt, 2011; Zoltai, 1993). The relatively dry peat at the surface provides insulation and inhibits downward energy transfer during the summer (Seppälä, 1986). Late fall rain events typical to the region (Prowse, 1990) increase moisture levels prior to freezing. When the peat freezes, the thermal conductivity increases and facilitates energy loss in the late fall and over winter. The process eventually leads to the aggradation of permafrost bulbs, and eventually allows permafrost to re-establish in the bog (Robinson and Moore, 2000; Zoltai, 1993). Observed climate warming in northern Canada appears to have slowed or halted the aggradation phase and favoured permafrost degradation (Quinton and Baltzer, 2013; Thie, 1974). In a climate favouring permafrost degradation, collapse-scar features appear to be growing rapidly and there is a growing need to understand the impact of these features on the basin water budget.

1.2 RESEARCH SITE:

All field work for this thesis was conducted in the Scotty Creek Research Basin in the southern Northwest Territories (NWT), Canada (Fig. 1). Background information about the study site is presented here, while detailed information can be found in the subsequent chapters. The SCRB is typical of other circumpolar high boreal peatlands. Permafrost exists exclusively underneath treed (primarily *Picea mariana*) permafrost cored peat plateaus and is interspersed by a mosaic of wetlands (channel fens and flat bogs) (Hayashi *et al*., 2004; Quinton *et al*., 2003). Here, a highly porous unsaturated layer provides thermal insulation to the permafrost from warm air temperatures in the summer. Further protection is provided by shading of a black spruce canopy, to limit the input of shortwave radiation (Chasmer *et al*., 2011). Within large permafrost plateau structures are collapse-scar fens and bogs (Robinson and Moore, 2000). Due to the volumetric expansion of ice in the permafrost core, plateaus typically rise about 1-2 m above the adjacent wetlands (Wright *et al*., 2008). This induces a hydraulic gradient between water stored in the suprapermafrost aquifer of the plateau and the surrounding wetlands. As a result, water is typically shed from peat plateaus to wetlands, making the plateaus a key producer of water input for the wetlands (Wright *et al*., 2008). During the thawed season, runoff from plateaus can be between 350 and 700 mm (Quinton and Baltzer, 2013; Wright *et al*., 2008). In comparison, basin runoff during the same time period (2003-2010) is significantly less (between 80 and 180 mm; WSC, 2016), illustrating the importance of plateaus as runoff producers.

During the spring and summer, runoff from hillslopes in permafrost terrains occurs in the thawed portion of the active layer. In peat-dominated terrains, the hydraulic conductivity of unfrozen soil decreases by 2-3 orders of magnitude within the top 20 cm (Quinton *et al*., 2008). During early spring and periods of high moisture input in the summer and fall, the water table occupies the zone of high transmissivity, enabling rapid subsurface runoff (Morison *et al*., 2016;

Wright *et al*., 2008). Seasonal thaw within the active layer exerts control on subsurface runoff and can exhibit substantial spatial variability, even at local scales (Wright *et al*., 2009). As permafrost thaws, the top of the permafrost layer retreats downward and increases the depth of the unfrozen layer capable of conducting flows of both mass (*i.e.* water) and energy.

Permafrost thaw along the southern edge of discontinuous permafrost has been well documented (Quinton *et al*., 2011; Jorgenson *et al*., 2010; Kwong and Gan, 1994). In the peatlands that characterise the landscape at Scotty Creek, thawing permafrost results in ground subsidence of the forested plateaus. As the ground subsides, the unsaturated layer that protects the permafrost disappears. This causes tree mortality through water logging of the root network (Baltzer *et al*., 2014), and also decreases the thermal offset between the permafrost table and ground surface (Smith and Riseborough, 2002) that had previously allowed permafrost to persist. As a result, the thawing of permafrost culminates in the transition of forested plateaus into wetlands, and a reduction of the areal extent of permafrost (Quinton *et al*., 2011). Increased fragmentation of permafrost plateaus has been found to cause more rapid permafrost thaw, as edge areas are more susceptible to thaw than the colder interiors of plateaus (Baltzer *et al*., 2014). The thaw of permafrost plateaus results in the expansion of permafrost-free wetlands. There are two dominant wetland types in the SCRB: channel fens and flat bogs. The hydrological function of each of these units is discussed in detail in chapters two and three.

1.3 RESEARCH OBJECTIVES:

Climate warming and associated permafrost thaw in northwestern Canada is occurring at an unprecedented recorded rate. Global climate models predict temperature increases to continue, which will have a direct impact on permafrost and hydrological conditions (Vincent *et al*., 2015). Wetland dominated catchments in discontinuous permafrost are common in the high boreal forest across the circumpolar north, yet few studies have attempted to quantify the impact of thawing permafrost on the basin water balance. This overall objective of this study is to evaluate how thawing permafrost will change hydrological connectivity and impact the lateral transfer of surface and subsurface water. This objective will be completed through the following sub-objectives:

1) Establish a causal link between increasing stream flows and the permafrost thaw-driven land cover changes in the Scotty Creek basin;
2) Investigate the mechanisms controlling the transmission of water through a series of cascading bogs within a permafrost cored peat plateau; and
3) Describe how small changes to the surface energy balance can induce talik growth and a peat plateau and examine the influence of taliks on the rate of permafrost thaw.

These sub-objectives will be assessed in each subsequent chapter of this thesis. Together, the mechanisms and processes described in the paper will enhance our ability to predict permafrost-thaw induced changes to the hydrological cycle in wetland dominated, discontinuous permafrost terrains. We hope to use the information presented here to guide predictive tools and models to determine the fate of freshwater resources in this region in the future.

1.4 THESIS OUTLINE:

Chapter 2:

Connon, RF, Quinton, WL, Craig, JR, and Hayashi, M, (2014). Changing hydrologic connectivity due to permafrost thaw in the lower Liard River valley, NWT, Canada. *Hydrological Processes*, 28: 4163-4178.

The first manuscript evaluates how recent permafrost thaw (1977-2010) is changing the runoff contributing area of basins in the lower Liard River valley in Canada’s Northwest Territories (NWT). Observed increases in basin runoff between 1996 and 2012 could not be explained by increased precipitation alone, leading to the theory that the basin is becoming more efficient at transporting water to the drainage network. Remotely sensed imagery was used to quantify the amount of forest loss (used as a proxy for permafrost loss) that occurred over a 6 km^2 area between 1977 and 2010. It was found that as wetland area increased (at the expense of permafrost plateaus), so too did the connectivity of the wetlands. Previous work (Quinton *et al*., 2003; Robinson and Moore, 2000) show that collapse scar features limit lateral transfer of surface and subsurface waters, as they are surrounded by relatively impermeable permafrost. Permafrost thaw allows these features to coalesce with each other and with the basin drainage network. As a result, the basin contributing area increases in response to permafrost thaw and can fundamentally change the hydrology of the region.

This manuscript also looks at other contributing factors that may cause runoff to increase (*i.e.* changes to precipitation and/or evaporation, direct contribution of meltwater from thawing permafrost, and increased baseflow from potential reactivation of groundwater pathways), and found that these were relatively insignificant compared to the changes in hydrologic connectivity. The concept of changing hydrologic connectivity had not previously been evaluated for a wetland dominated northern boreal forest, and we demonstrate that these changing processes can yield substantial changes to the basin water budget.

When this paper was written in 2014, the most recent available runoff data was from 2012. As an addendum, it should be noted that in the three subsequent years (2013 to 2015), total basin runoff has been very low in response to three very dry years. Similar to ideas presented by

Hewlett and Hibbert (1967), we hypothesize that contributing areas expand and contract dependent on moisture conditions. The increased hydrologic connectivity documented here is the result of bog features becoming more connected to the channel fen network. Ingram (1978) shows that the hydraulic conductivity in the catotelm (dark, well decomposed peat with small pore size) can be orders of magnitude smaller than the acrotelm (light, relatively undecomposed peat with large pore size). It can be reasoned that when precipitation is low, the water table in the bogs resides close to the catotelm, limiting transport through bogs and effectively disconnecting them from the drainage network. The observed changes to hydrologic connectivity are therefore most pronounced during periods of high precipitation input, when the water table is at, or near the ground surface and in the highly conductive layers of peat. These concepts are further evaluated in chapter three of this thesis.

Chapter 3:

Connon, RF, Quinton, WL, Craig, JR, Hanisch, J, Sonnentag, O. (2015). The hydrology of interconnected bog complexes in discontinuous permafrost terrains, *Hydrological Processes,* 29: 3831-3847

The first paper in this thesis evaluated changing hydrologic connectivity at a landscape scale and relied on visual changes available from aerial photographs and remotely sensed imagery. Expanding on those ideas, the second paper uses field observations to quantify the interconnectivity of collapse scar bogs. Using a combination of IKONOS and Landsat imagery, Quinton *et al*. (2003) suggest that bogs at Scotty Creek are isolated from the drainage network and are effective water storage areas. Field observations show that in some cases drainage channels may form between collapse scar bogs and may allow large areas of peat plateau complexes to drain to the fen network. The relatively narrow (< 5 m) drainage channels were not observable in the IKONOS (pixel size: 4 m x 4 m) analysis in Quinton *et al*. (2003) and ground-based measurements were used to document their extent. The drainage channels form a hydrological connection between the bogs, forming a 'bog cascade', whereby the bog furthest upstream drains into a downstream bog, and so-forth, eventually discharging into the channel fen. The aim of this study is to quantify the amount of runoff produced from these complexes and examine their influence on the basin hydrograph.

This study found that runoff from bog cascades can vary spatially and is highly dependent on the plateau:bog ratio, as plateaus contribute large amounts of runoff to the bogs. The amount of runoff is also dependent on the storage levels within the bogs, similar to 'fill-and-spill' concepts developed in the Canadian Shield (Spence and Woo, 2003). If antecedent moisture levels are low, the bogs are not capable of transporting water through the drainage channels and act as storage reservoirs. However, when water levels are high (such as during the spring freshet), water is readily transported through the drainage channels and the runoff contributing area of the basin is much greater. These dynamic changes to runoff contributing areas are likely to change as permafrost thaw continues and allows more drainage channels to form.

Chapter 4:

Connon, RF, Quinton, WL, Devoie, É, Veness, T, Hayashi, M. (in review). The influence of shallow taliks on permafrost thaw and active layer dynamics in subarctic Canada. *Journal of Geophysical Research – Earth Surface*

Quantifying permafrost loss using remotely sensed imagery is relatively easy in areas such as Scotty Creek, as forest loss can be used as a proxy for permafrost loss (Chasmer *et al*., 2011; Jorgenson *et al*., 2001). This method allows for detection of permafrost loss dating back to the earliest availability of aerial photographs (1947 for Scotty Creek). Physical ground-based measurements of permafrost are more difficult due to the scarcity of long term datasets in northern regions. Measurements of suprapermafrost layer thickness and permafrost width have been taken at a single plateau at Scotty Creek since 1999 (see Quinton and Baltzer, 2013). Thaw rates at this plateau have been very high; width has decreased by approximately 1 m $year^{-1}$ and suprapermafrost layer thickness has increased at a rate of over 0.05 m $year^{-1}$. This chapter expands on this research and looks at permafrost thaw rates at nine transects over a five year period (2011 to 2016) and describes mechanisms that may be associated with more rapid permafrost thaw.

It was found that permafrost thaw is not consistent throughout the basin. Some areas thawed much more rapidly than others, as they were therefore undergoing land cover change at a much faster rate. The major finding from this chapter was that in some areas of Scotty Creek the thaw depth exceeds the depth of re-freeze the following winter. This results in the formation of a shallow talik between the active layer and the permafrost. This finding is important, as many studies (including CALM) assume complete re-freeze, as measurements of active layer thickness are only made at the end of summer and do not include a measure of re-freeze.

This chapter investigates the significance of the talik layer from a thermal perspective and postulates significance from a hydrological perspective. Energetically, permafrost thaw is dominated by latent heat requirements and sensible heat exchange can be considered negligible (Hayashi *et al*., 2007). The most rapid vertical permafrost thaw was observed at sites with taliks. There are two types of taliks found at Scotty Creek: connected taliks and isolated taliks. Connected taliks occur where a talik network fully connects two wetlands through an intervening plateau, while isolated taliks exist solely within a plateau. Connected taliks also offer a hydrologic pathway for water to be routed across a plateau year-round. This change to hydrologic connectivity has not been previously analyzed and could increase the availability of water to the basin drainage network.

The second finding of this chapter was how the prominence of taliks increased over four years. The first year of measurements occurred in 2011, where 20% of points had a talik. In 2016, this number increased to 48%. We hypothesize that an anomalous year with high temperatures and increased ground heat flux (leading to deep thaw) followed by a winter with high early season snow accumulation (as occurred in 2012-2013 at Scotty Creek) can produce a talik network that is likely irreversible in the current climate. If the areal coverage of a talik network increases, and areas with taliks experience more rapid thaw, permafrost thaw is likely to display a non-linear response to climate warming. This chapter elucidates how talik development influences permafrost thaw, and when viewed from the context of chapters two and three, increased permafrost thaw can fundamentally change the basin water budget.

CHAPTER 2:
Changing hydrologic connectivity due to permafrost thaw in the lower Liard River valley, NWT, Canada

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2.1 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 (Dery *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*., 2005). Déry and Wood (2005) reported that rivers flowing into Hudson Bay and the Labrador Sea experienced declines in discharge between 1964 and 2003; however Dery *et al*. (2009) show a reversal of this trend when expanding the observational period by just four years (1964 – 2007). This exemplifies the high inter-annual variability of stream flow records and also illustrates the influence of high stream flows in recent years on long term trends. Reasons for increasing stream flow 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 stream flow in permafrost covered basins. One example is additional moisture inputs 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 (Rowland *et al*., 2010; St. Jacques and Sauchyn, 2009, Walvoord and Striegl, 2007).

St. Jacques and Sauchyn (2009) found that nine of 23 rivers analyzed 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 wide-spread 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, since 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 (McCylmont *et al*., 2013). Permafrost in this southern margin is also warm (> -0.5 °C) and relatively thin (<10m) (Burgess and Smith 2000), compared to 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 since the surface of the permafrost bodies rise above adjacent permafrost-free terrains, thereby impounding or re-directing 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 1947-2008 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.

2.2 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 2-1). Permafrost extent in this region is sporadic discontinuous (~40% coverage) and has high ice content 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 (Aylsworth 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 record (1996-2012), while annual adjusted precipitation (Mekis and Vincent, 2011) at Fort Simpson is 540 mm over that same period (47% snow, 53% rain). Hydrometric data from the four basins show an apparent increase in runoff patterns over the last 15 years (Figure 2-2); while 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 sub-surface 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 2-3) 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 2-4) 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 has not yet been studied.

Evapotranspiration (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 (March 29 – June 4) and 2005 (April 19 – June 8), 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 due to higher air 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, due to 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, since the plateau flanks and corresponding hydraulic gradient (average gradient of 0.049; n: 15; SD: 0.025 (Chasmer *et al*., 2011)) slope toward 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, since 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).

2.3 METHODOLOGY:

2.3.1 Permafrost thaw

Since, in the type of system studied here, permafrost only exists underneath treed peat plateaus (Quinton *et al*., 2011), 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 digitized 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 area of interest (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 (LiDAR) imagery (up to 10 returns per square meter) 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 recognized 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 normalized 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, based on the average rate of thaw as calculated over the 6 km^2 AOI.

2.3.2 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 2-1 for location). Due to 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 (October 1 to September 30) instead of the calendar year, as snow accumulation beginning in October is not recognized 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 2-1), 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.

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.* Dery 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 interannual variations. Figure 2-5 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.

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 January 01 to March 31 (Walvoord and Striegl, 2007; St. Jacques and Sauchyn, 2009). These values were converted to mm using the same method as section 3.2. This yielded the cumulative depth (mm) of winter baseflow expected during the 3 month period from January 01 to March 31. 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.

2.3.3 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 stream flow.

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 per cent 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), while 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 based on each basin's proximity to the other (Figure 2-1).

The low standard deviations suggest that it is valid to estimate runoff ratios for Scotty Creek and Blackstone River for the period before hydrometric data was 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 hypothesized 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 above to provide estimates for the increases in runoff resulting from the direct increase in area of wetlands connected to the drainage network as well as the increase 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 stream flow-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 to the observed increase calculated as the difference between the 1996 and 2012 annual runoff values on the Kendall-Theil robust line for each basin.

2.4 RESULTS AND DISCUSSION:

2.4.1 Increases to Stream flow

An increase in annual total runoff began in 1996 for all basins (Figure 2-5). 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 1996-2012 (annual increases of 8.3 and 9.4 mm yr^{-1} respectively). Scotty Creek ($p \leq 0.01$) and Blackstone River ($p \leq 0.05$) also show significant trends from 1996-2012 with annual increases of 8.5 and 6.6 mm yr^{-1} respectively. Precipitation increased significantly ($p \leq 0.1$) over the period of 1996-2012 at a rate of 5.7 mm yr^{-1}; however did not show a significant change from 1976-1995 (Table 2-2). 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 2-2). Runoff ratios also increased significantly ($p \leq 0.1$) in all four basins since 1996 (Table 2-2).

2.4.2 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 causes wetlands to coalesce, it appears that storage capacity in bogs is decreasing while 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 2-6c,d and is termed ‘bog capture’.

Analysis of remotely sensed imagery and aerial photographs shows that permafrost covered 3.27 ± 0.3 km^2 in 1977 and decreased to 2.85 ± 0.23 km^2 by 2008 over the 6 km^2 AOI (Table 2-3). This represents a 13% (±2.7%) loss of permafrost coverage between 1977 and 2010. As described above, permafrost loss transforms forested area to wetland, and since 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 non-linear response to warming. As outlined in Figure 2-4, 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 higher 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 2-4 shows the depth of runoff that would be 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, while 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 runoff after accounting for additional inputs (Table 2-4). Quinton *et al*. (2003) show that the per cent cover of flat bog at Scotty Creek is 10.2% compared to 3.4% (Table 2-1) 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 2-7 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.

2.4.3 Changes in Runoff Contributing Area

Growth of the secondary plateau contributing area is difficult to quantify from aerial photographs and remotely sensed imagery due to uncertainties of drainage divides on peat plateaus. Due to 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 (see Figure 2-4) 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*. (2011) 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 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 hypothesized 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 recognized 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 2-4).

2.4.4 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 2-2) 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 is 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). Due to the relatively low 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 (Wu *et al*., 2005; Manabe *et al*., 2004). 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 2-8). Precipitation 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 is 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) as well as the second (2012), third (1999), fourth (2007) and sixth (2005) largest recorded total annual snowfalls for the 1898-2012 period of record (Table 2-5). There is a possibility that the apparent increase in stream flow was driven in part by these several large annual snowfalls 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*., (2013) 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, due to 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 2-8).

2.4.5 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 yr^{-1} (Table 2-6). As initial ice content is lowered, the amount of additional moisture input decreases by just over 0.1 mm yr^{-1} (Table 2-6). This amount is minimal over the short term; however, it may be of more importance over longer time 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 2-4. 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 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.

2.4.6 Additional Groundwater Contributions

The literature suggests that permafrost thaw is reactivating groundwater pathways (Walvoord and Striegl, 2007) and thereby increasing baseflow and subsequently stream flow rates (St. Jacques and Sauchyn, 2009). Using winter flows as a proxy to estimate changes to baseflow, it is shown that baseflow contributions appear to be minimal ($\leq$ 7% of total annual runoff); however each basin shows a 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 yr^{-1} in the four basins. Annual baseflow contributions to total annual stream flow range from 0.6% (Scotty) to 6.1% (Jean-Marie) based on 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.

2.5 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 on-going 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 non-linear 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.

2.6 FIGURES

Figure 1

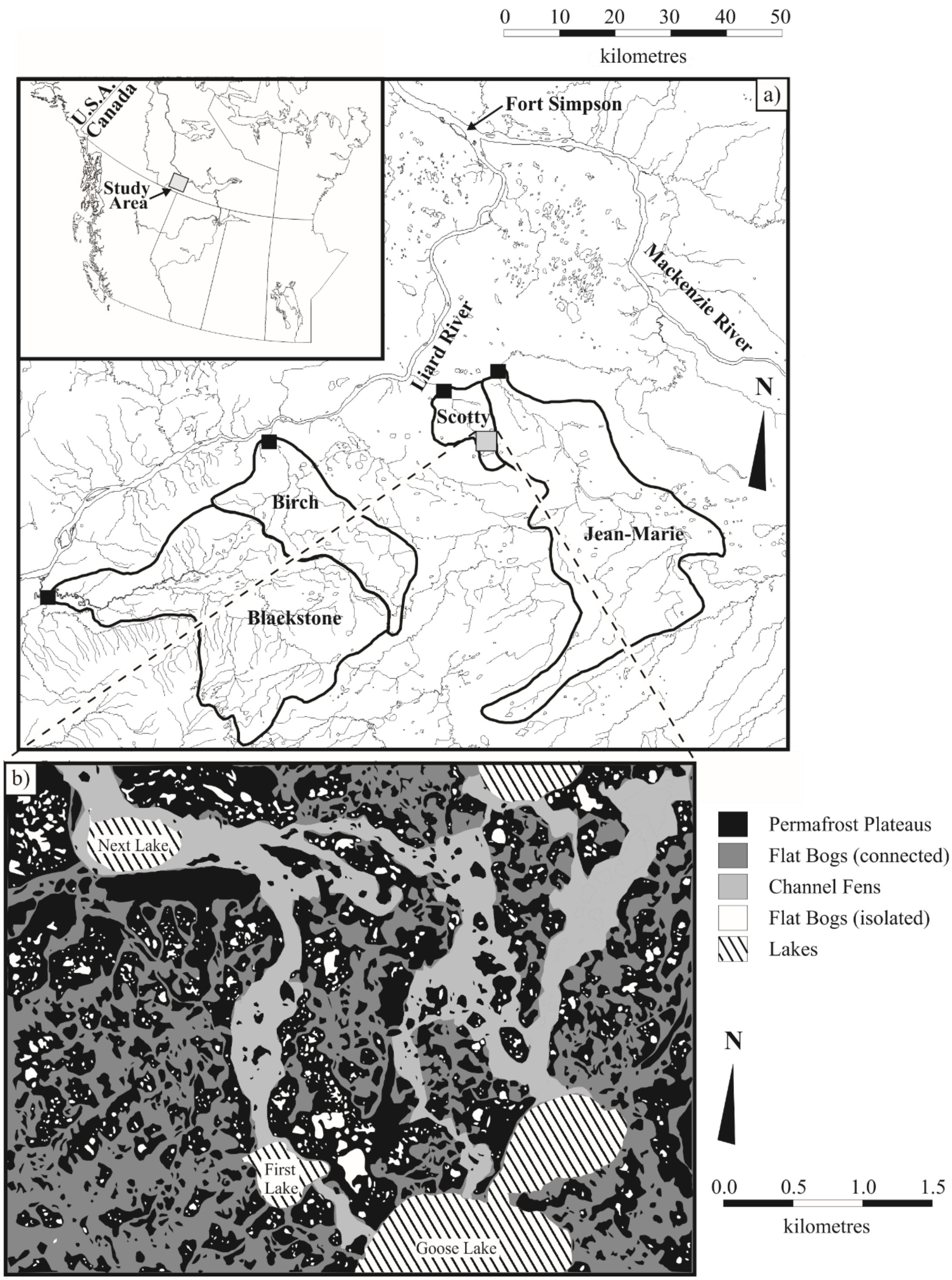

Figure 2-1: (a) Location of the four study basins in the Lower Liard River Valley: Blackstone River (2050 km^2); Birch River (542 km^2); Scotty Creek (152 km^2); and Jean Marie River (1310 km^2). (b) Inset presents a classified, 22 km^2 image of the Scotty Creek basin outlining the major land-cover types.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

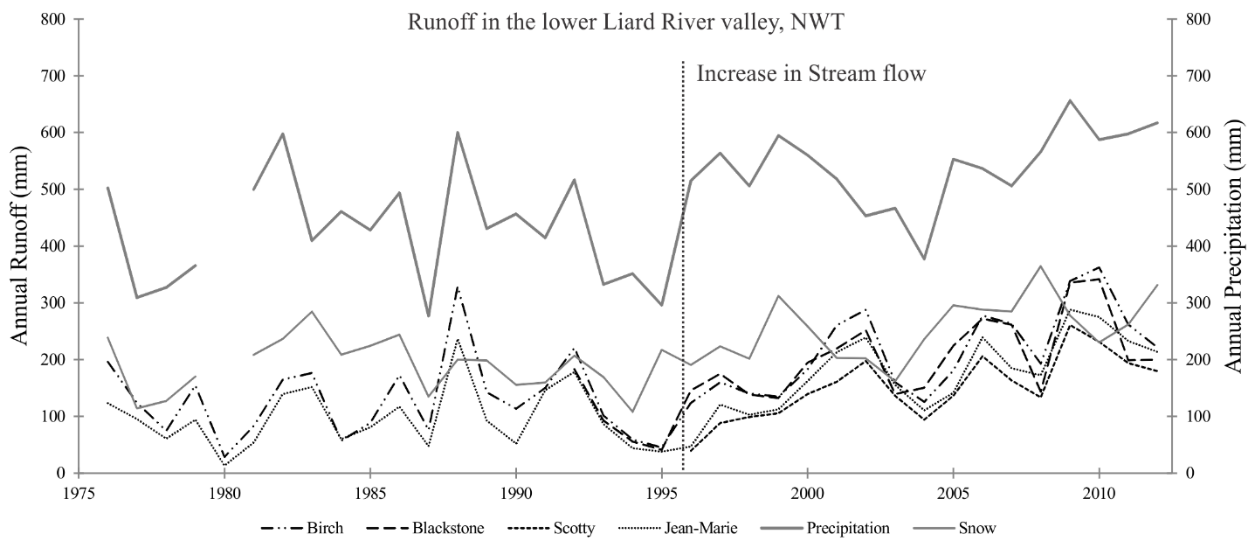

Figure 2-2: Total annual runoff at Birch River, Blackstone River, Scotty Creek and Jean-Marie River and total annual precipitation (*i.e.* rain and snow) at Fort Simpson for the period 1975 to 2012. Both values were computed for the water year Oct. 01 – Sep. 30.

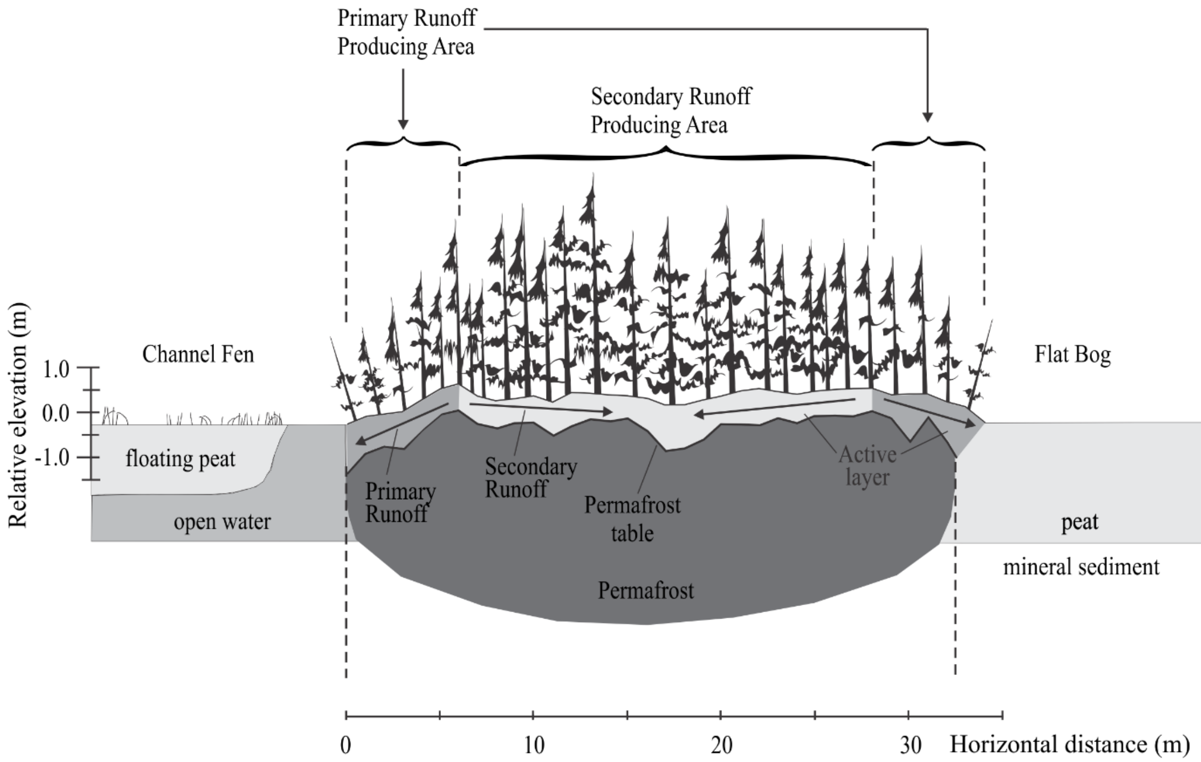

Figure 2-3: Cross sectional area of a peat plateau underlain by permafrost with a channel fen and flat bog on either side. Primary runoff is shown to occur on the flanks of the peat plateau, flowing into adjacent wetlands. Secondary runoff flows into internal bogs on the peat plateau. Note the depression 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.

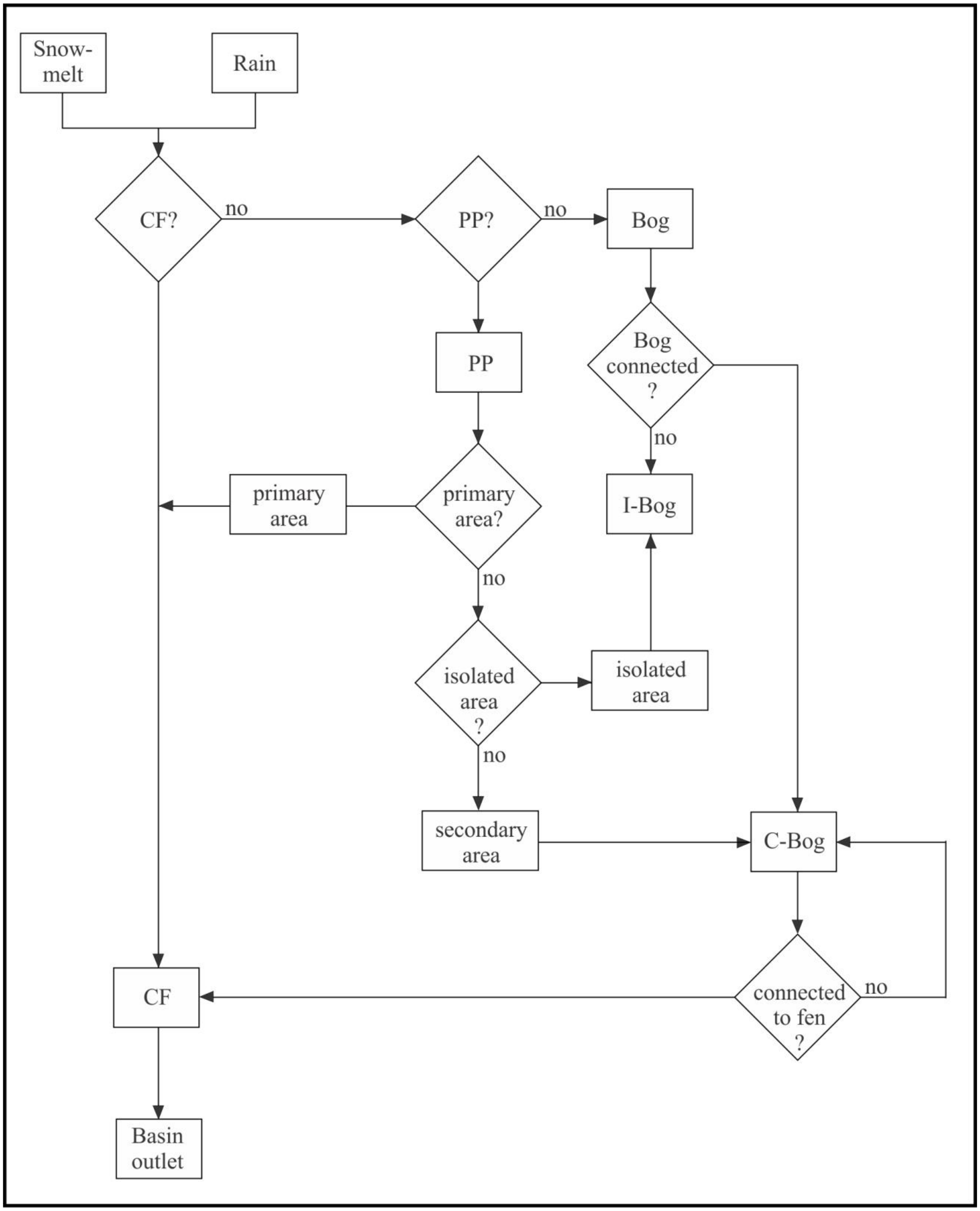

Figure 2-4: Conceptual model of pathways for precipitation in a wetland-dominated basin in the zone of discontinuous permafrost. Rectangles represent processes; diamonds represent questions. Primary runoff flows from a peat plateau (PP) directly into the channel fen (CF) whereas secondary runoff is routed 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).

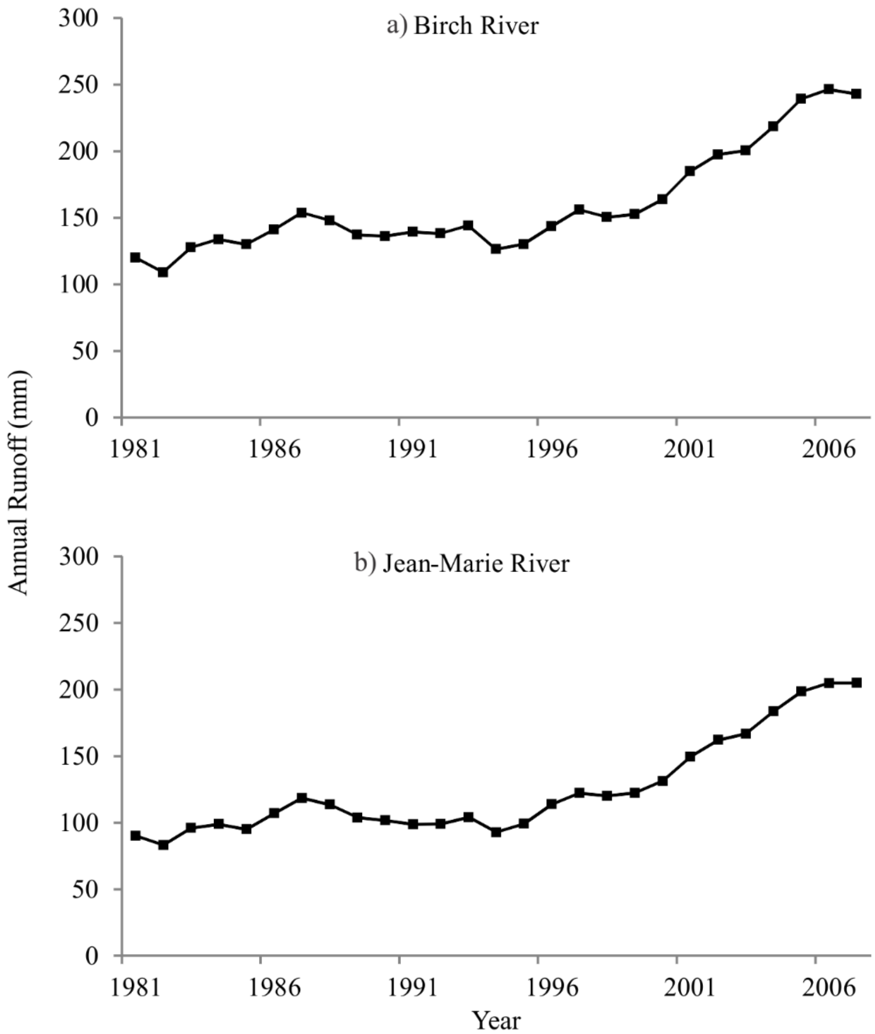

Figure 2-5: 11 year moving windows of annual runoff at a) Birch River and b) Jean Marie River. The year indicated is the median year (*i.e.* 1990 represents the period from 1985 to 1995).

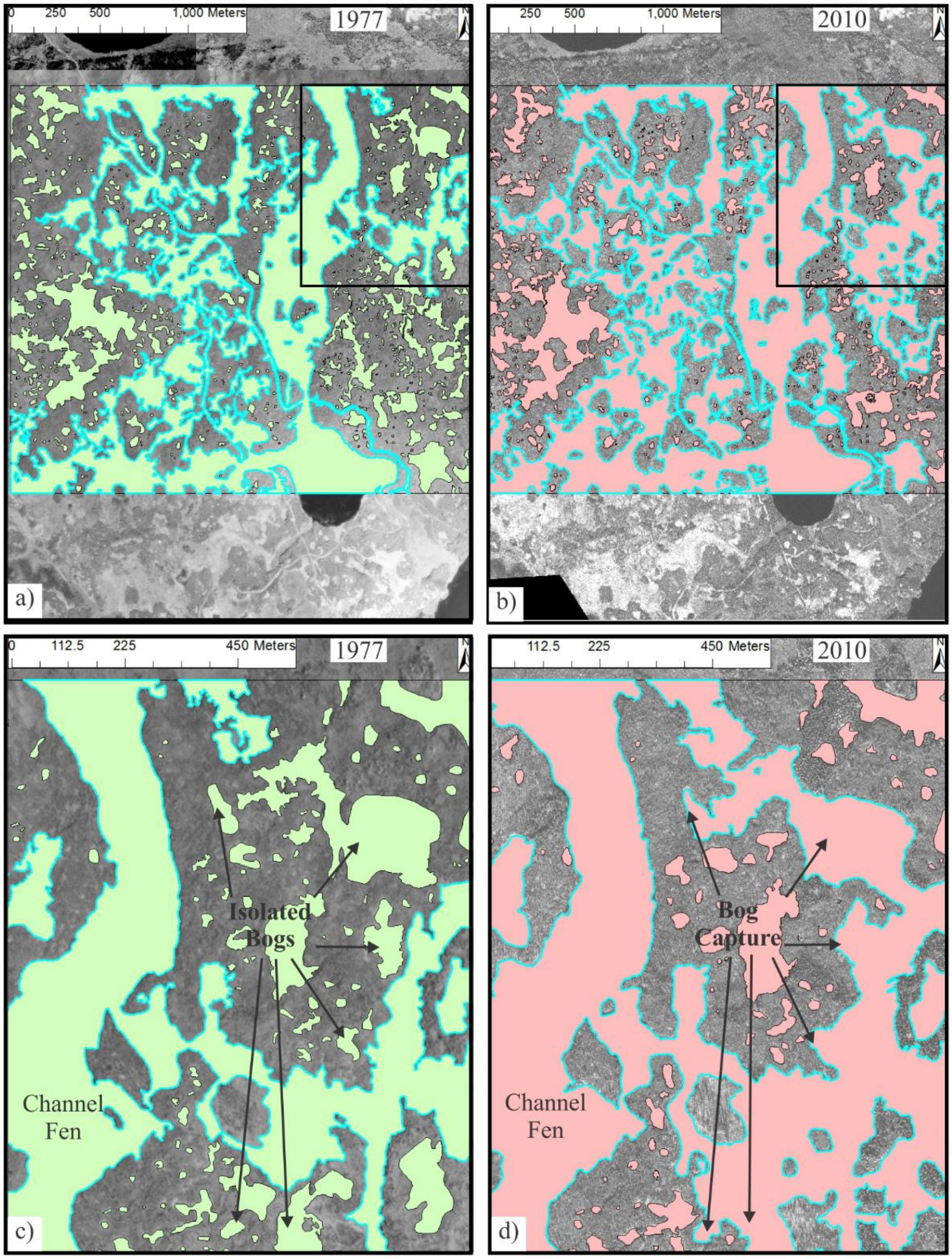

Figure 2-6: 6 km^2 area of interest (AOI) at Scotty Creek basin for (a) 1977 and (b) 2010 outlining permafrost and permafrost free areas. Permafrost-free areas (*i.e.* wetlands) are shaded in green (1977) and red (2010). Sections of wetland outlined in blue represent area that is connected to the basin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

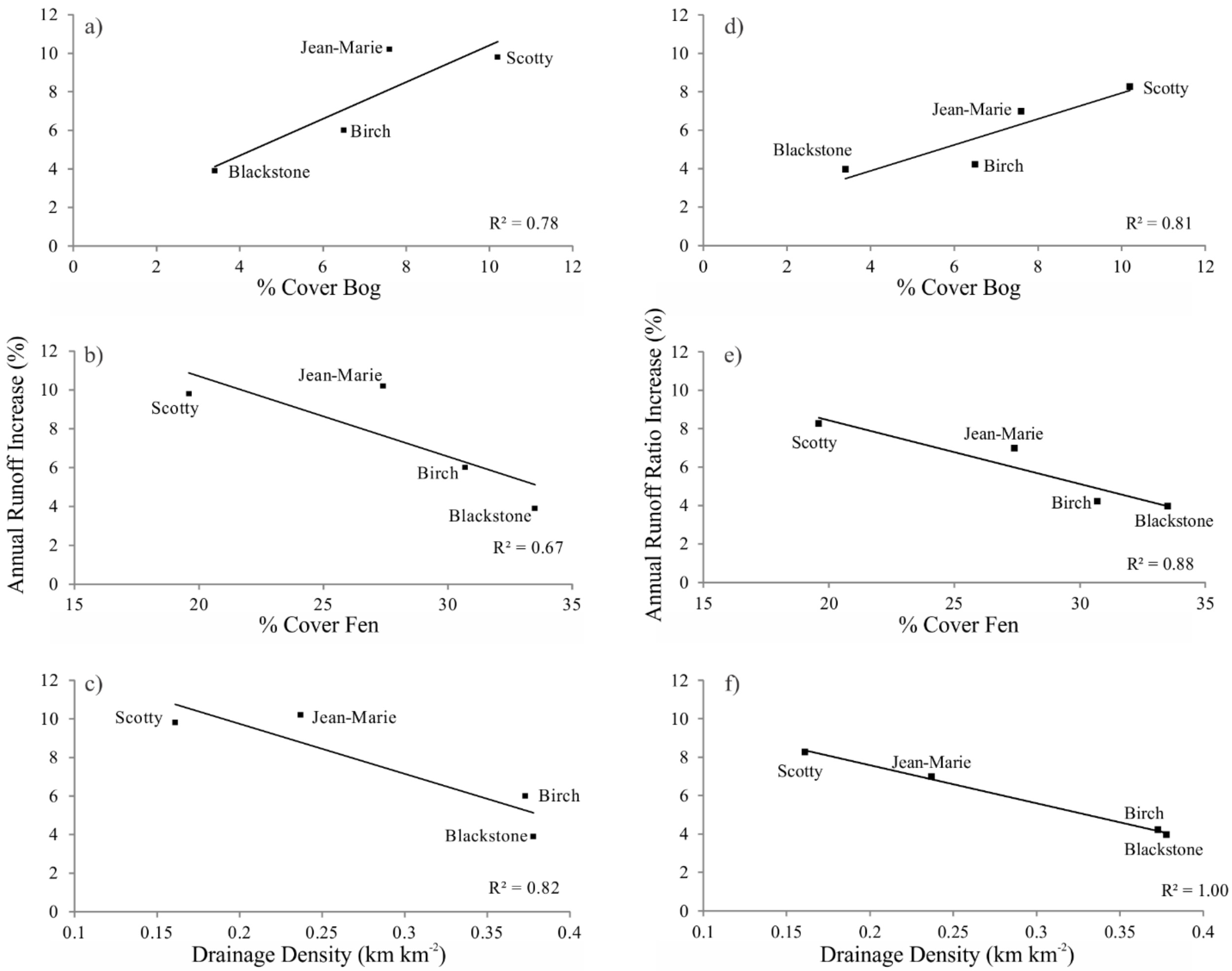

Figure 2-7: Annual increase (%) in runoff and runoff ratios plotted against: (a,d) per cent cover of bog; (b,e) per cent cover of fen; and (c,f) drainage density for four basins in the lower Liard River valley.

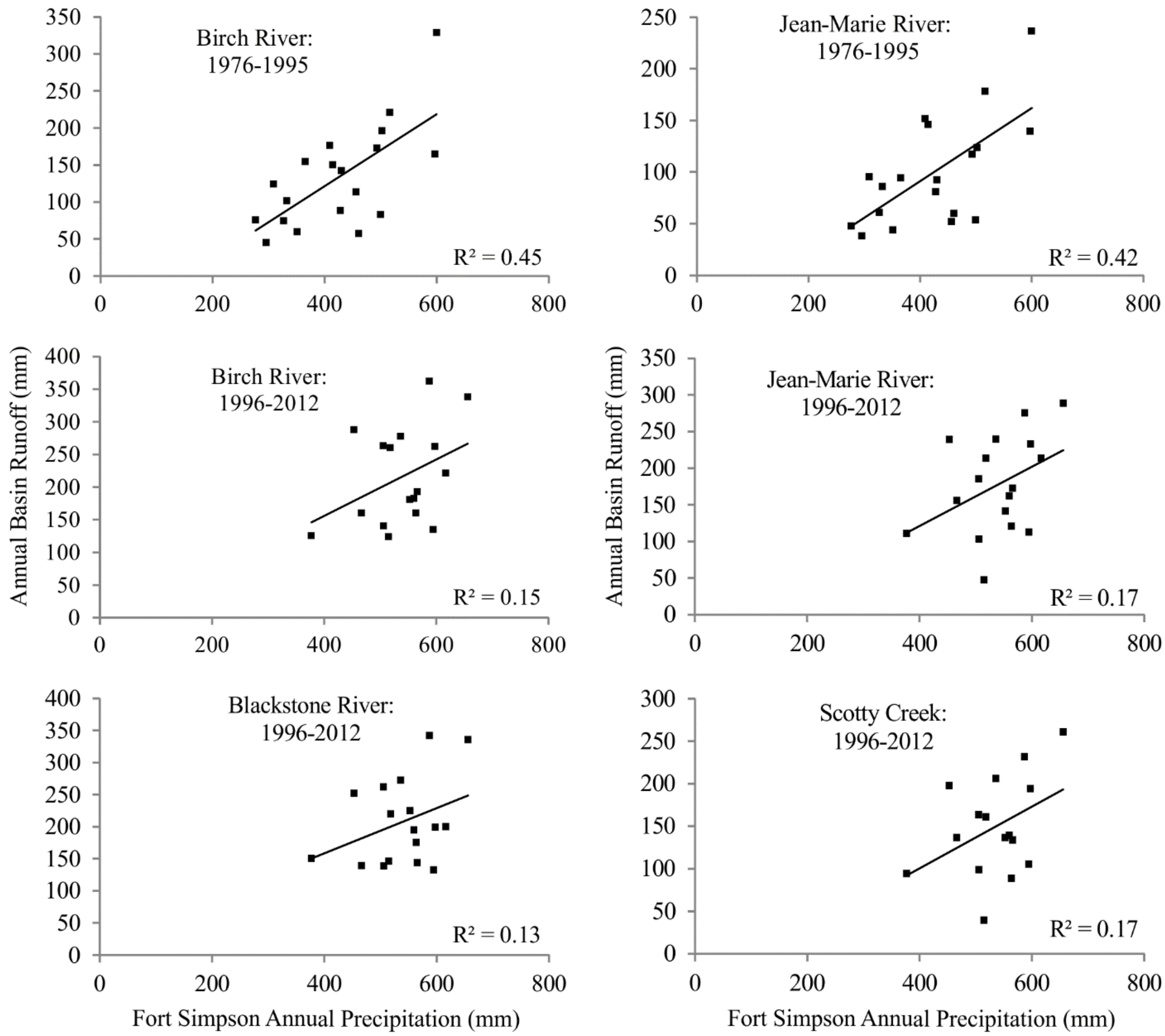

Figure 2-8: Plots of annual basin runoff against annual precipitation for two time periods (1976-1995 and 1996-2012) for four basins in the lower Liard River valley. For the period of 1976-1995, annual basin runoff is significantly correlated to annual precipitation for both Birch River and Jean-Marie River ($p<0.01$).

2.7 TABLES:

Table 2-1: Area and per cent cover of each land-cover type at the four basins in the lower Liard River valley. Adapted from Quinton *et al.*, (2003).

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

Table 2-2: 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). Precipitation values were obtained at Fort Simpson, NWT over the period of 1976 to 2012. Mean change (% and mm) are derived from the Kendall-Theil robust lines. Italicized, bolded and underlined values represent statistical significance at $p \leq 0.1$, $p \leq 0.05$, $p \leq 0.01$ respectively.

	Annual Runoff: 1976-1995					Runoff Ratios: 1996-2012			
Basin	Average Runoff (mm)	Total Change (mm)	Average change/yr (mm)	Average change/yr (%)	Basin	Average Runoff Ratio	Total Change	Average change/yr	Average change/yr (%)
Birch	127.8	-22.9	-1.1	-0.9	Birch	0.400	***0.188***	***0.011***	***4.2***
Jean-Marie	95.5	-12.4	-0.6	-0.7	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***

	Annual Runoff: 1996-2012				Annual Baseflow: 1996-2012				
Basin	Average Runoff (mm)	Total Change (mm)	Average change/yr (mm)	Average change/yr (%)	Average Baseflow (mm)	Total Change (mm)	Average change/yr (mm)	Average change/yr (%)	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*

	Fort Simpson Annual Precipitation			
Period	Average Precip (mm)	Total Change (mm)	Average change/yr (mm)	Average change/yr (%)
1976-1995	424.7	-43.0	-2.3	-0.5
1996-2012	539.7	*97.2*	*5.7*	*1.1*

Table 2-3: 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%**

Table 2-4: Expected runoff values for the four basins in the lower Liard River valley calculated with the 25 per cent trimmed mean runoff ratios from 1976-1995 (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'). 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).

Basin	2012 Precipitation with 1976-1995 Runoff Ratios (mm)	+	-	Additional Annual Inputs (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**

Table 2-5: Ten highest years for corrected precipitation (total, snow and rain) data from 1898-2012 (Corrected as per 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

Table 2-6: The annual additional moisture inputs from the conversion of ice to water resulting from permafrost thaw for selected years. Permafrost depth and porosity are held constant at 10 m and 0.8 respectively. Area coverage is based on the Scotty Creek basin.

Year	Area Covered by Permafrost (%)	Area Covered by Permafrost (m^2)	Area of Permafrost Lost (m^2)	Volume of Ice in Basin (m^3)	Annual Additional Moisture Inputs ($mm\ yr^{-1}$)
1977	54.6	8.30×10^7	1.79×10^5	1.43×10^6	8.6
1980	54.0	8.20×10^7	1.77×10^5	1.42×10^6	8.5
1990	51.8	7.87×10^7	1.70×10^5	1.36×10^6	8.1
2000	49.6	7.55×10^7	1.63×10^5	1.30×10^6	7.8
2010	47.5	7.22×10^7	1.56×10^5	1.25×10^6	7.5

CHAPTER 3:
The hydrology of interconnected bog complexes in discontinuous permafrost terrains

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3.1 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; IPCC, 2014). 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 (Jones and Rinehart, 2010; St. Jacques and Sauchyn, 2012). 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 3-1), 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 permafrost bodies with high ice content, 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 *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 behavior 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) characterise 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.

3.2 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 3-1). 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 -2.8°C and average unadjusted precipitation is 390 mm yr^{-1} with 62% falling as rain and 38% as snow (MSC, 2014) between 1981 and 2011. The precipitation values reported here (as well as Chapter 4) are not from the adjusted and homogenized dataset as presented in Chapter 2. The reader is referred to Chapter 5, section 5.3.2 for a more detailed rationale to the differences in precipitation values between chapters. 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*., 2008). 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*., 2015). 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 (Zoltai, 1993; Robinson and Moore, 2000). These features are dominated by *Sphagnum* mosses (*i.e. Sphagnum fuscum, S.*

angustifolium and *S. riparium*) (Zoltai *et al*., 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 3-2, illustrating the different flow paths which are dependent on moisture conditions at the time of the input. Figure 3-3 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.

3.2.1 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-neighbor 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 due to the presence of permafrost surrounding the bogs and clay-rich glacial till with low hydraulic conductivity [$K_S = 1x10^{-9} – 1x10^{-10}$ 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 3-3). The average cross-sectional area of surface water in the channels during the spring freshet is 1.0 m^2 (n: 12; standard deviation: 0.5) with a range of 0.4 m^2 to 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 3-4 depicts the cross section 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 3-4c).

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 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).

3.3 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 3-3). 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 3-4b). The base of the weirs and flumes were installed between 0.5 m – 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 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 (2010) 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, 2010), 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, 2010). 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). Due to noise in the data, the east cascade hydrograph has been smoothed with a 3 hour 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 using a calibrated spring scale. 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 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 3-3). 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:

$$(1)\ ET = \alpha \frac{1}{\lambda}\left[\frac{s \cdot (Q^{*} - Q_{g})}{s + \gamma}\right]$$

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 $^{\circ}$C^{-1}), γ is the psychrometric constant (kPa

$^{\circ}C^{-1}$), 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) at 1.5 m above the ground surface. Q_g was measured using a ground heat flux plate (HFT3, Campbell Scientific) 0.05 m below the ground surface. Heat storage in the upper 0.05 m was added to the heat flux to obtain a measure of the ground heat flux at the ground surface (Mayocchi and Bristow, 1995). Air temperature was measured using an HMP45C probe (Campbell Scientific) 1.5 m above the ground surface, 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 in Bog W-3 in May of 2014 at a height of 1.5 m above the ground surface (Helbig *et al*., 2016). 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:

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

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.

3.4 RESULTS:

Results from two 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 3-5).

3.4.1 Channel Development

Figure 3-6 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 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}).

3.4.2Runoff Produced from Secondary Contributing Areas

Discharge from 2013 was estimated solely from manual measurements (Figure 3-7) 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 3-7b), 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 3-8). 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; however 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.

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.

3.4.3 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 3-1 shows the total discharge and runoff for the two bog cascades for two 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 due to drier conditions. The east cascade produced 52 mm of runoff, while the west cascade produced 6 mm.

Total snow accumulation was very different between the two 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. Average precipitation at Fort Simpson between 1981 and 2010 was 390 mm yr^{-1}. Both 2013 and 2014 received less than average precipitation (319 mm and 272 mm respectively), while the years 2009 to 2011 were wetter than average (519 mm, 505 mm, and 424 mm yr^{-1}). This allows for a comparison of soil moisture levels between wet and dry periods (Figure 3-9). 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 June 06 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 3-1). 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 3-5 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 3-5 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 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.

3.5 DISCUSSION:

3.5.1 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*. (2010) 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.

3.5.2 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, p 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 3-2).

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 due to dynamic and heterogeneous contributing areas. Large bogs with corresponding large storage capacities and small contributing areas (Table 3-2) 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-3 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 resulted in very low flow levels (Figure 3-9). 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 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 3-5). 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).

3.5.3 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 x 4 m) and Landsat (resolution: 30 m x 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 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.

3.6 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 meters. 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 programs are required to provide a better understanding of these systems.

3.7 FIGURES

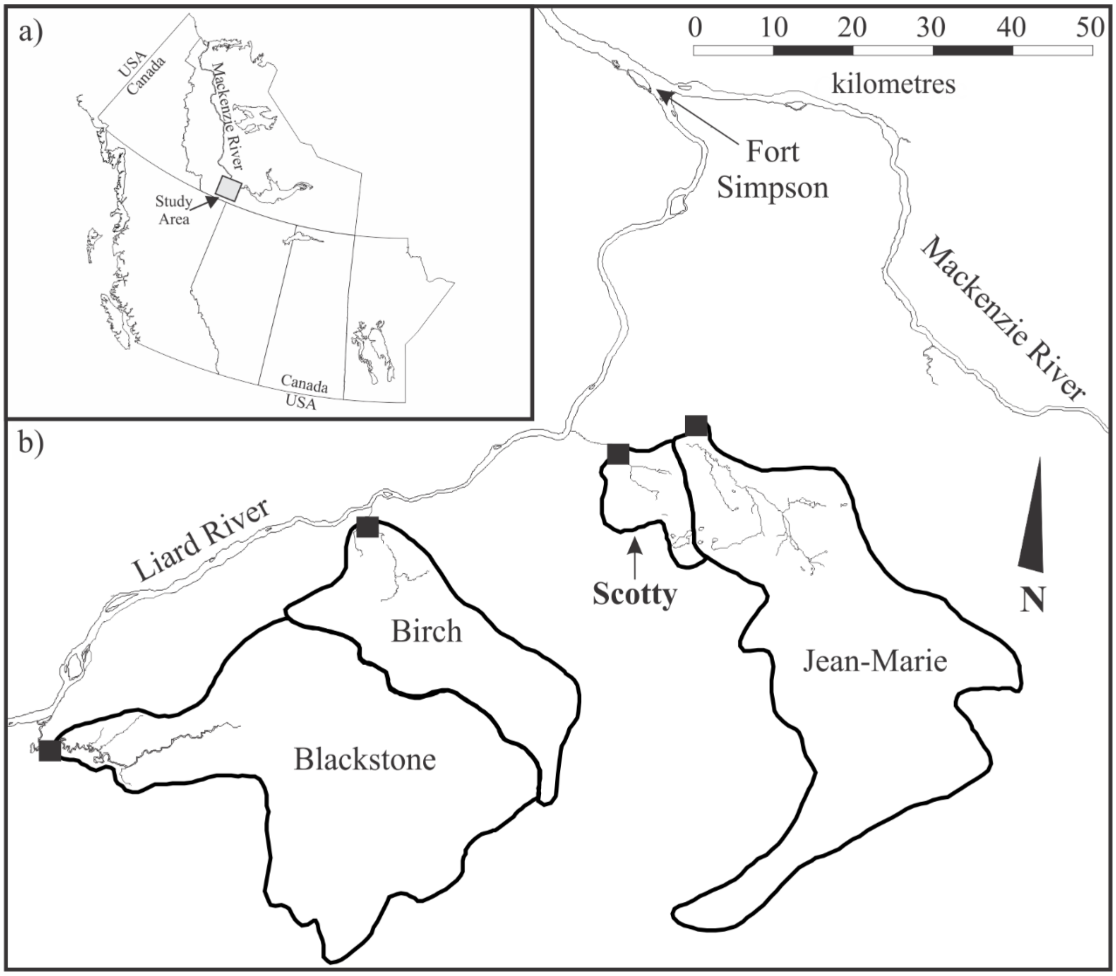

Figure 3-1: a) Location of the lower Liard River valley in the Northwest Territories, Canada; b) Inset of the Scotty Creek Research Basin. Black squares indicate Water Survey of Canada gauging stations and delineated basins represent the area upstream of the gauging station (not entire basin).

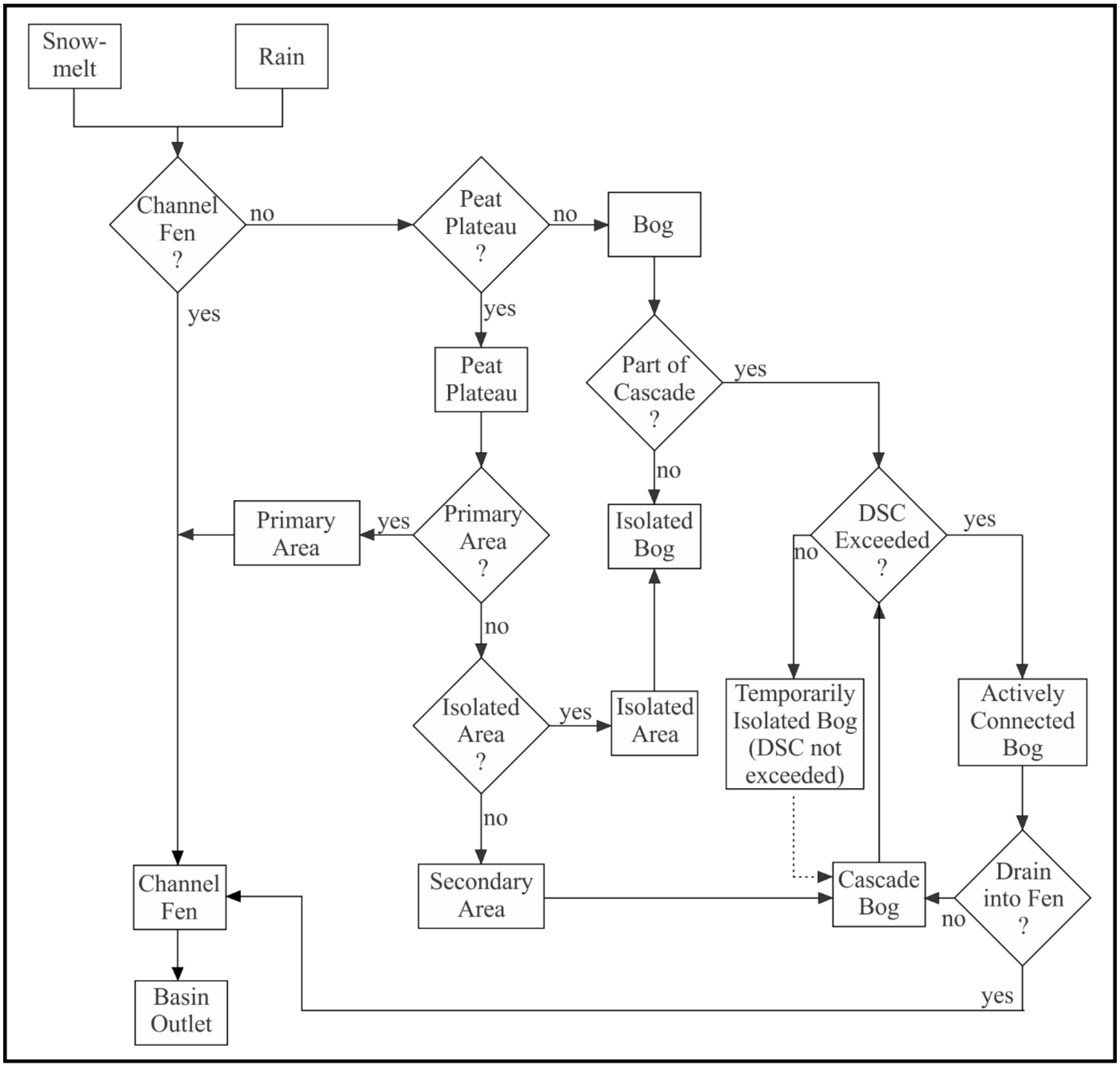

Figure 3-2: Conceptual diagram of the partitioning of precipitation inputs. The rectangles represent processes while diamonds represent questions. DSC: Depression Storage Capacity. Note that temporarily isolated bogs may reconnect with the rest of the cascade during rain events. Neglecting to include 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.

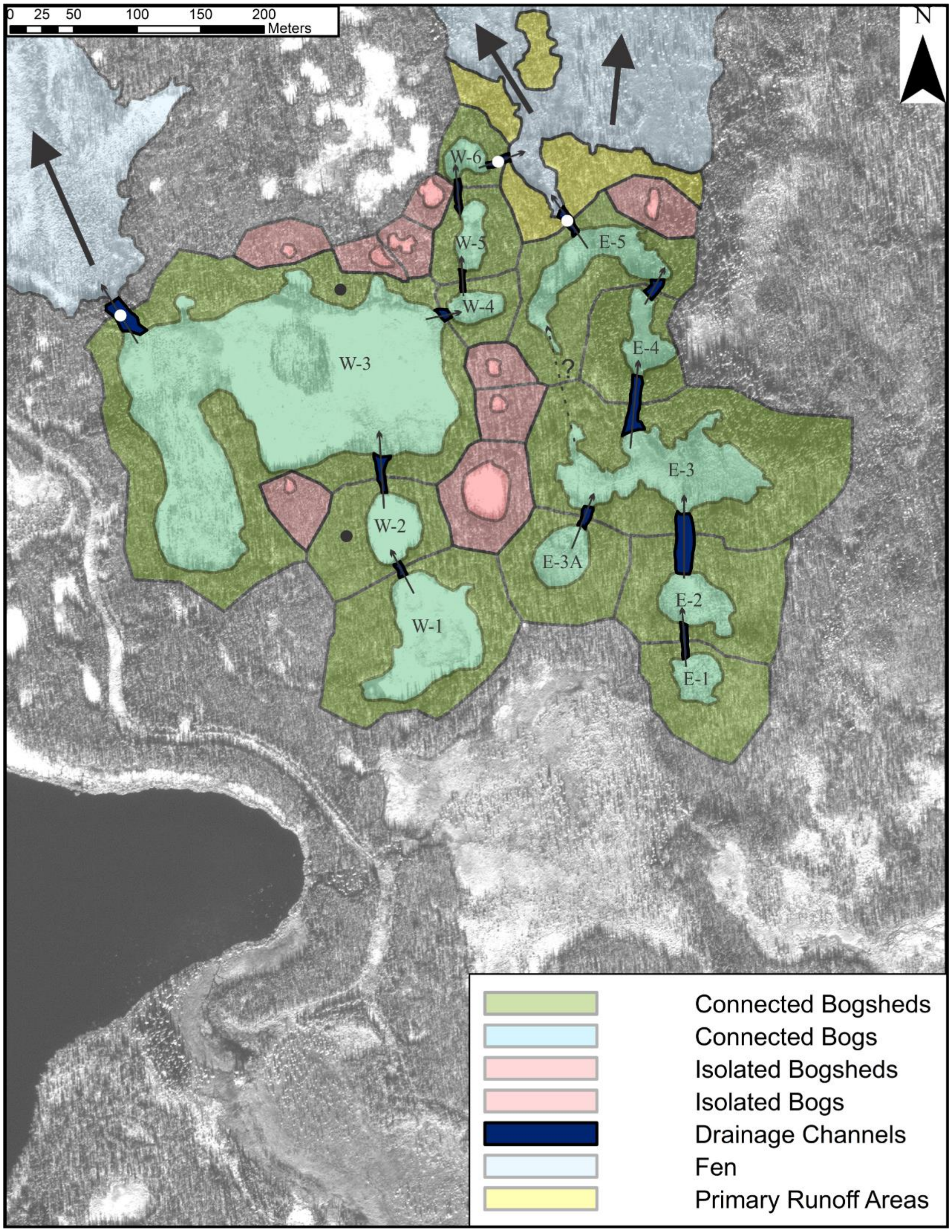

Figure 3-3: Map of two gauged bog series cascades at the Scotty Creek Research Basin. Note that bog W-3 has two outflows (W-3A to the northwest and another to bog W-4). Due to extremely flat topography in bogs, the drainage divide in bog W-3 is unknown. Black arrows indicate flow direction; white dots 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.

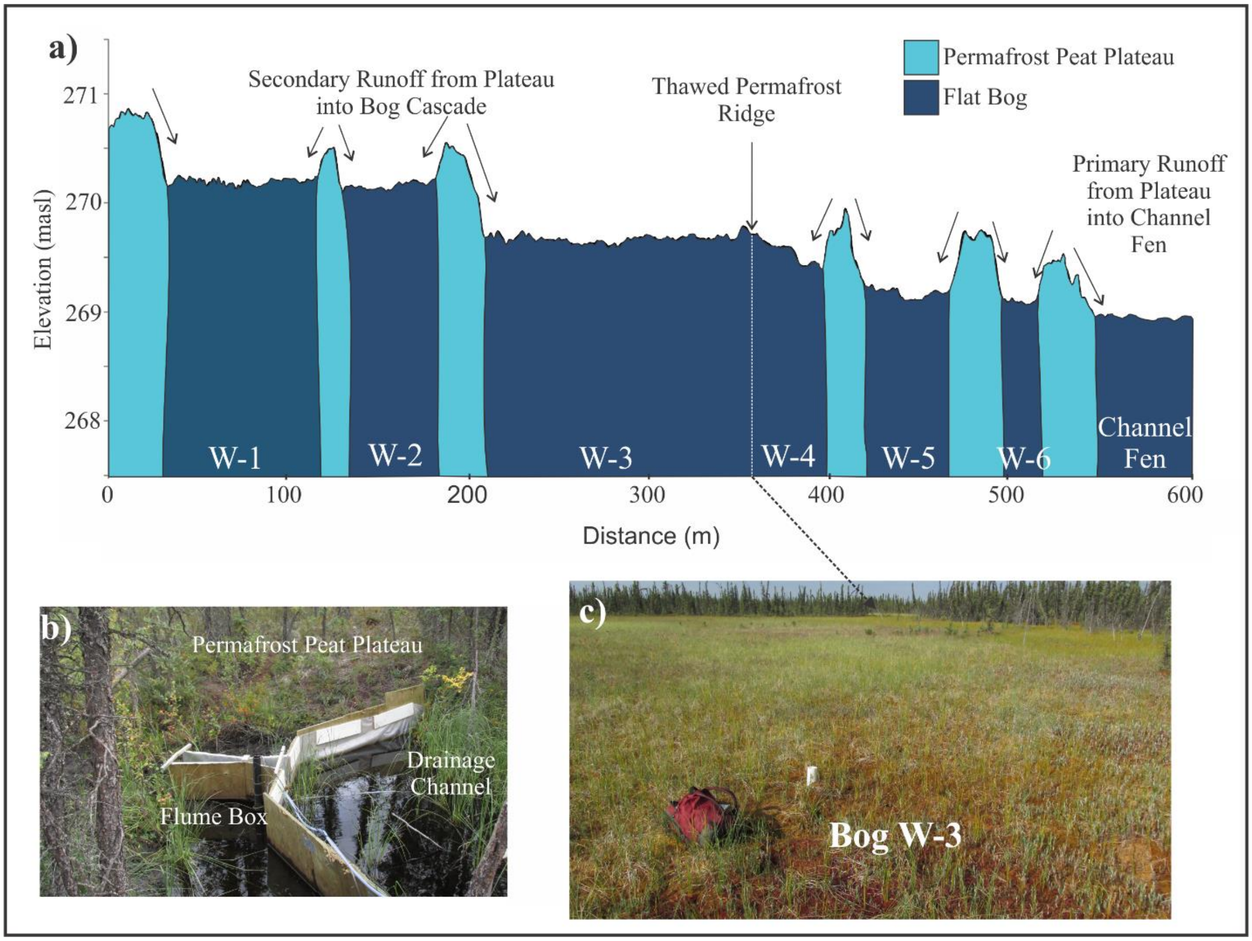

Figure 3-4: a) Cross sectional area of bog cascade. Actual elevations were derived from a digital elevation model (DEM). The permafrost ridges encircle each bog with the exception of the drainage channels (not illustrated in diagram); b) Photograph of a drainage channel and a flume box used for gauging 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.

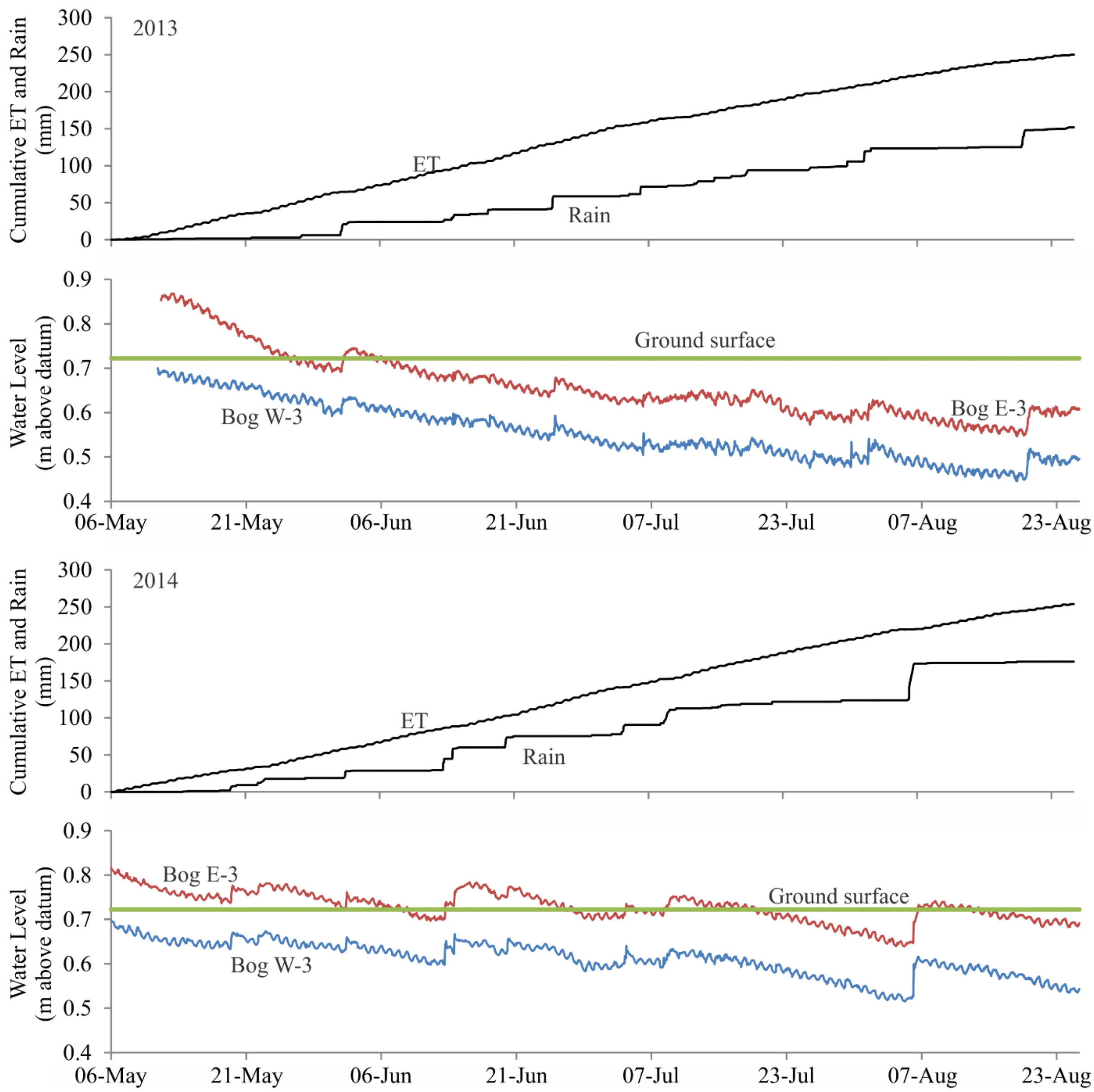

Figure 3-5: Cumulative evapotranspiration (ET) and rain as well as water table levels for bogs W-3 and E-3 in 2013 and 2014. The water table elevation is relative to an arbitrary datum and is shown in relation to the ground surface.

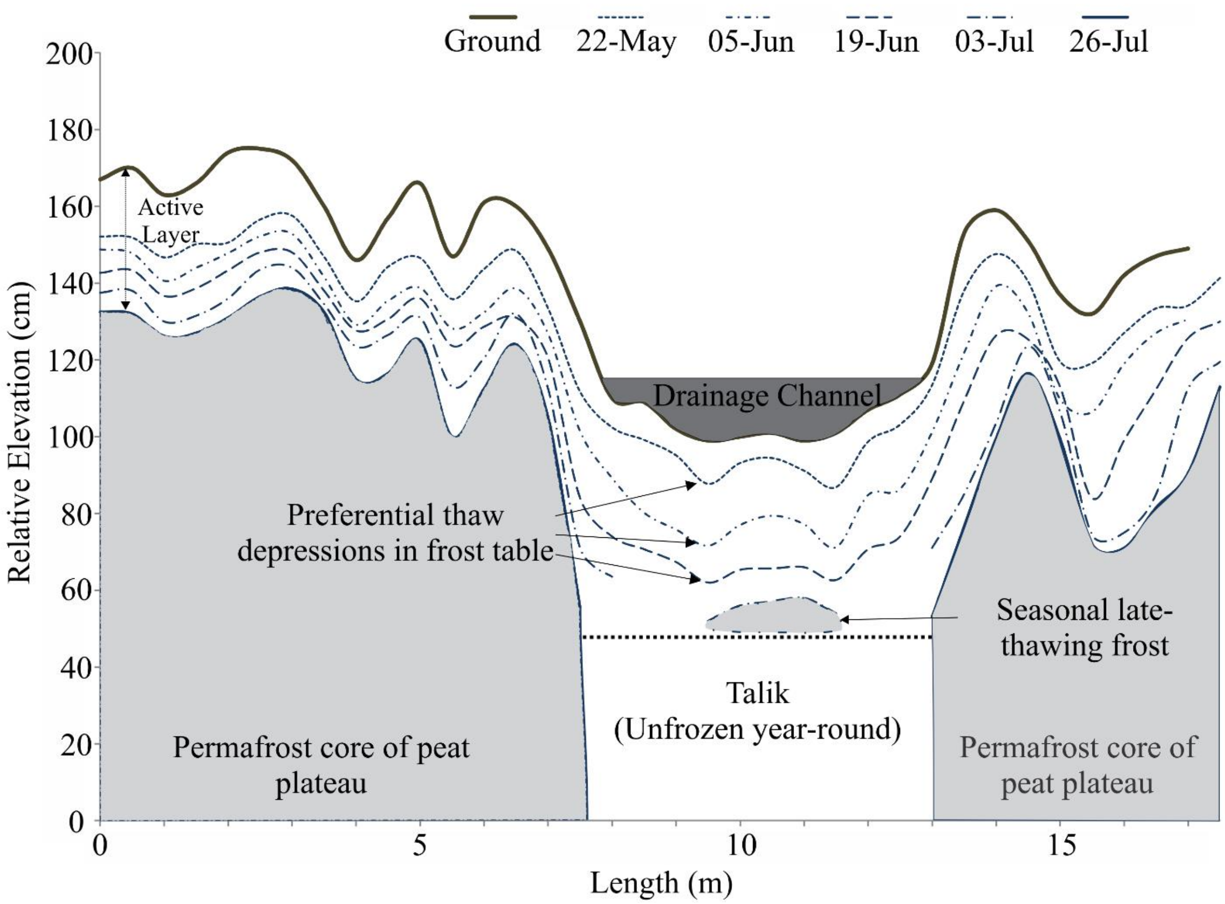

Figure 3-6: A cross section of a drainage channel (W-6). The dashed lines indicate the thawing of the frost table over the course of the spring/summer. Surface water exists primarily during the spring freshet when all bogs are connected. As the frost table lowers due to seasonal thaw, so too does the water table perched above it. Note that there is a 5x vertical exaggeration to denote the preferential thaw depressions in the frost table.

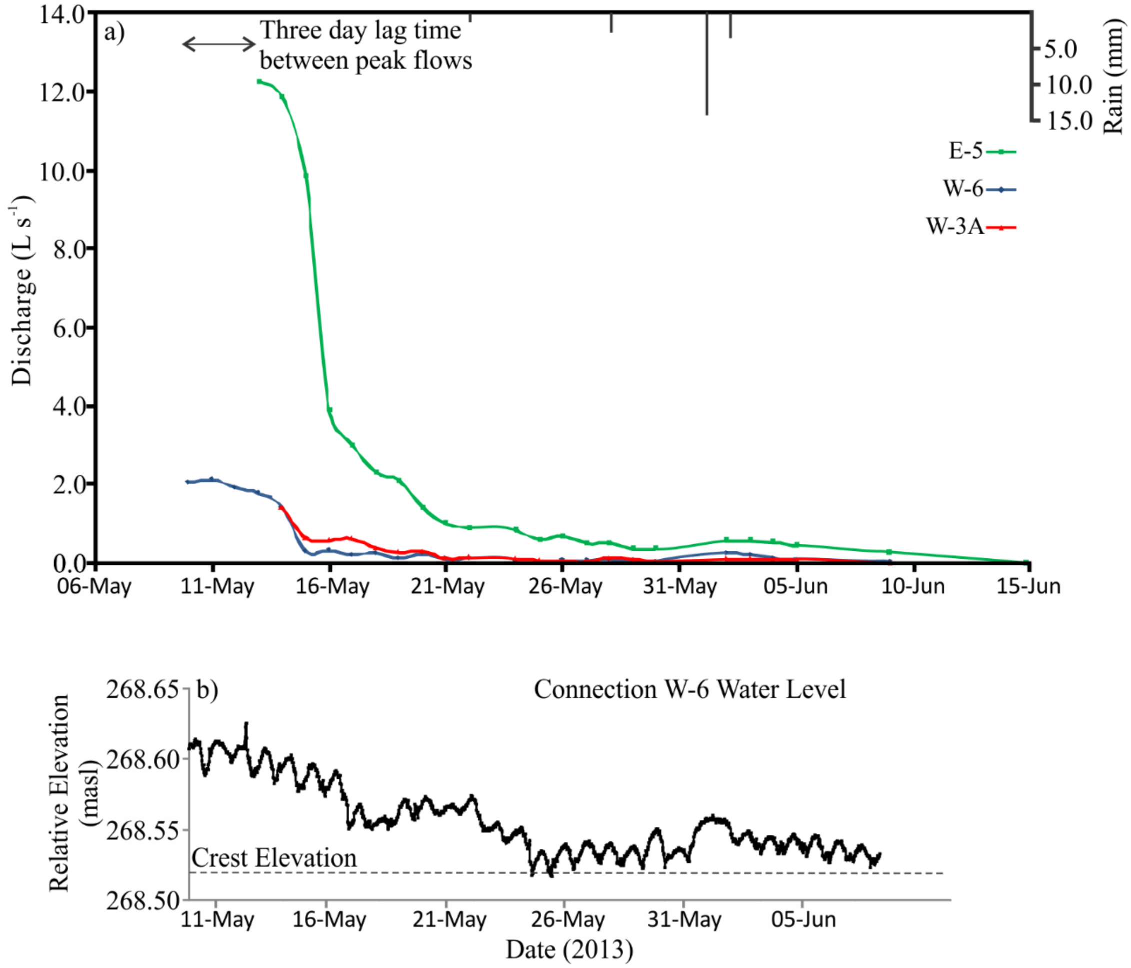

Figure 3-7: a) Discharge from the three terminal drainage channels in 2013. Channel E-5 was snow-choked until peak flows began May 13; b) Water level records from channel W-6. The diurnal fluctuations are thought to be an artifact of the barometric pressure transducer being exposed to ambient air temperature 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.

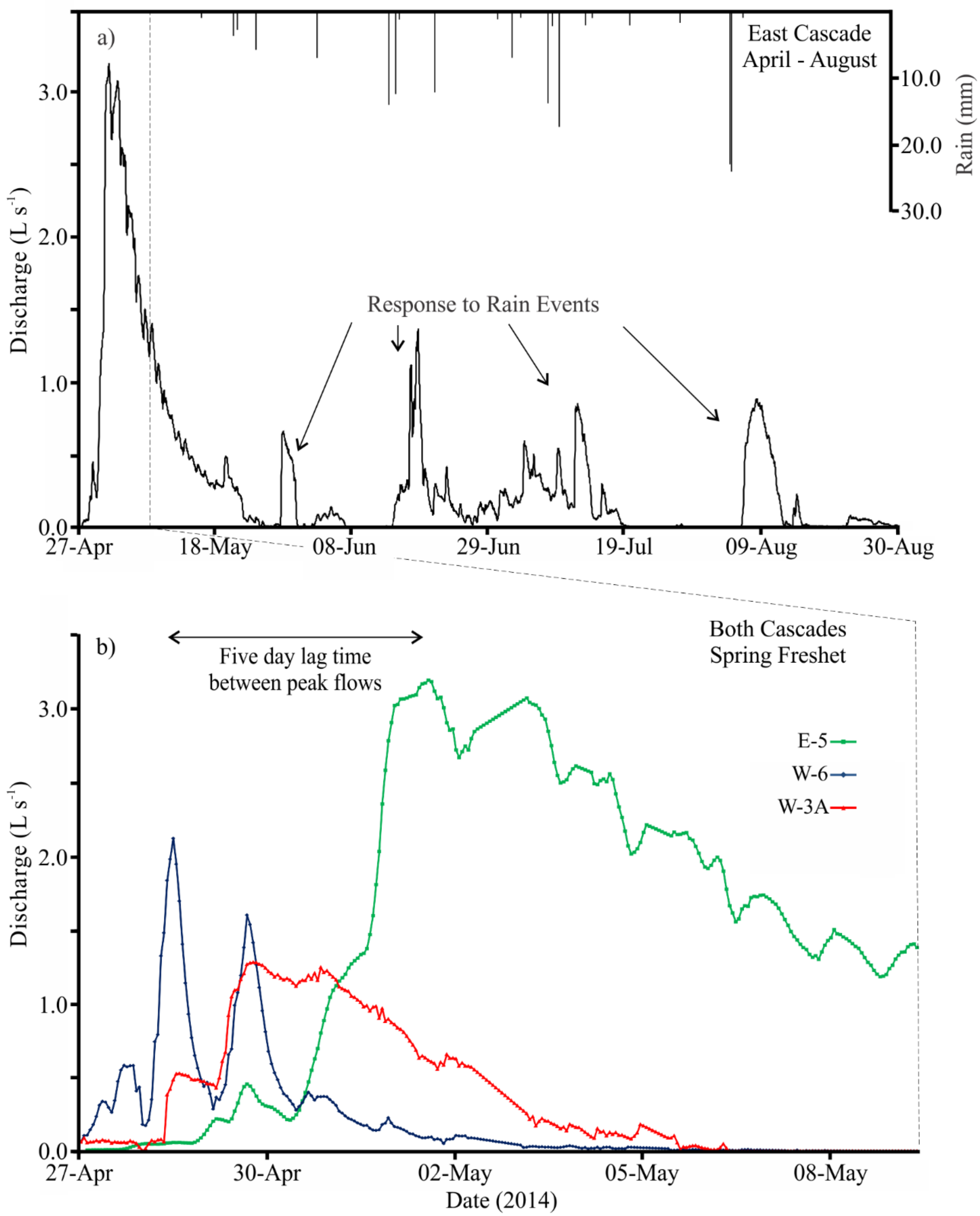

Figure 3-8: Total discharge in 2014. a) Hydrograph from the east cascade for the entire season. Snowmelt dominates the hydrograph, however response to rain events is evident; b) Hydrographs for the three terminal connections during snowmelt, 2014 (no substantial rain events during this period). The peaks from channel W-6 occur around 19:00 each day and are likely a response to daily snowmelt.

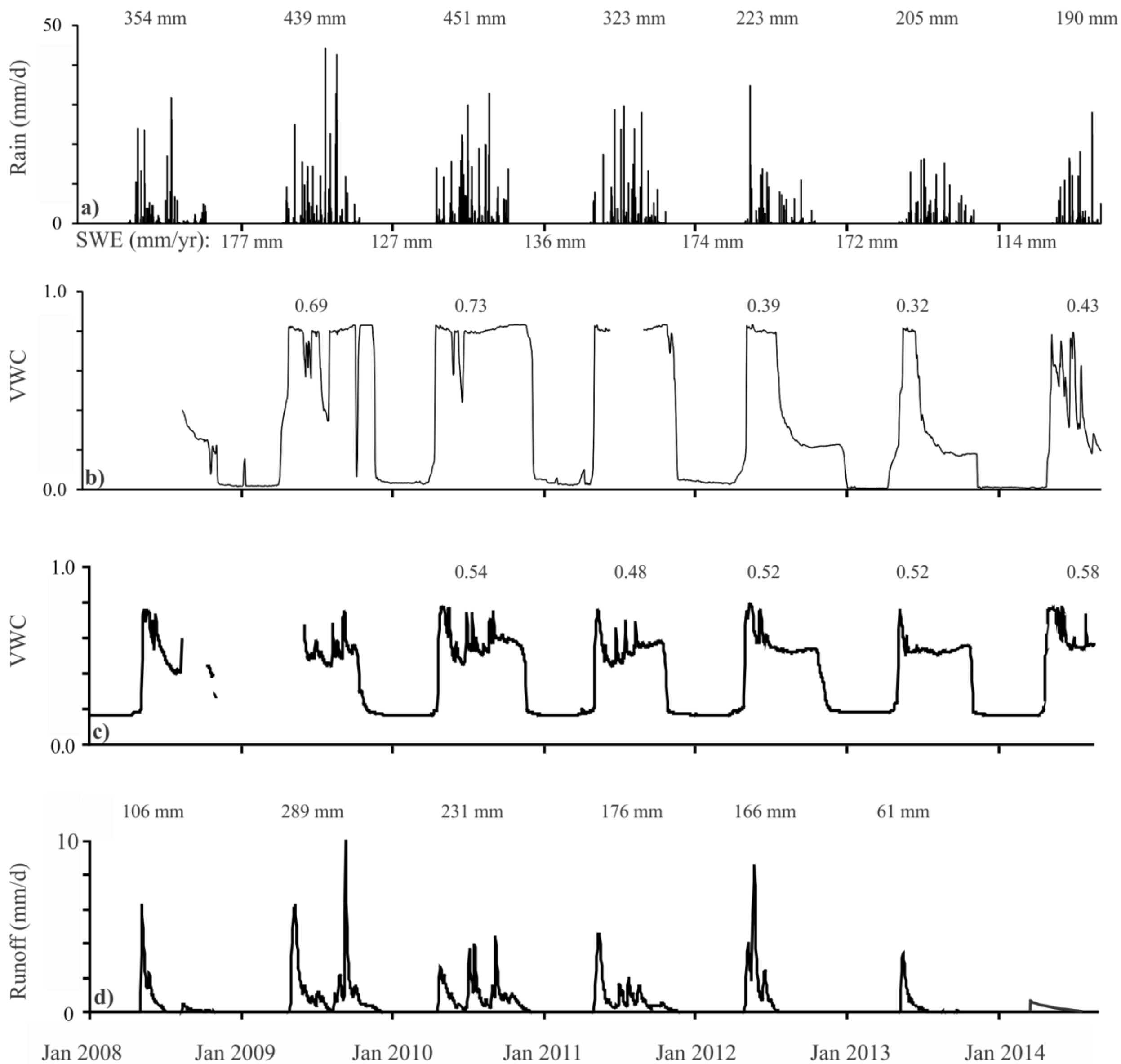

Figure 3-9: Precipitation (a), soil moisture (b, c) and runoff (d) data from 2008 to 2014 at Scotty Creek. Total rainfall, snow water equivalent (SWE) and runoff values are shown on each graph, while average soil moisture values for the thawed season are given. SWE values for each season are indicated 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.

3.8 TABLES

Table 3-1: Total precipitation, discharge and runoff ratios for snowmelt and rain events in 2013 and 2014.

Year	Cascade	Event (Date)	Total Discharge		SWE	Rain	Runoff Ratio
			(m^3)	(mm)	(mm)	(mm)	
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
		09-Jul-14	133	2.0	0	16.5	0.12
		06-Aug-14	309	4.5	0	46.5	0.10
		Season	3519	51.7	115	158.8	0.19

Table 3-2: 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	34421	953	1069	829	45056	1240	2058	7099	1730	1616	3503	17246
Plateau area (m^2)	9862	5438	22092	1930	3259	1901	44483	5818	8687	18509	5798	4481	7586	50879
Bogshed area (m^2)	15681	7404	56513	2883	4328	2730	89539	7058	10745	25608	7528	6097	11089	68125
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^2)	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)	18155	5819	101886	2134	1881	1891	131766	1934	2799	8519	2837	1681	5885	23655

CHAPTER 4:
The influence of shallow taliks on permafrost thaw and active layer dynamics in subarctic Canada

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4.1 INTRODUCTION:

The active layer is the top layer of ground subject to annual thawing and freezing in areas underlain by permafrost [*ACGR*, 1988]. In permafrost terrains, most ecological and hydrological processes of interest occur in the active layer, and consequently, there is a need to measure and document changes to active layer thickness (ALT) over long periods of time [*Tarnocai et al*., 2004; *Brown et al*., 2008; *Bonnaventure and Lamoureaux*, 2013]. The thickness of the active layer is dependent on the surface energy balance, which is affected by many factors, including air temperature [*Kane et al.*, 1991; *Åkerman and Johansson*, 2008; *Sannel et al*., 2016], snow accumulation and soil moisture [*Guglielman*, 2006; *Johansson et al*., 2013; *Atchley et al*., 2016], vegetation composition [*Jean and Payette*, 2014], slope aspect [*Carey and Woo*, 1999] and angle [*Hannell*, 1973]. Natural or anthropogenic changes to these variables can lead to changes in ALT.

Continental permafrost exists at high latitudes and altitudes, and typically where the mean annual air temperature (MAAT) is below 0°C. MAAT has increased significantly in northern Canada over the last 65 years [*Vincent et al*., 2015], and IPCC AR5 climate models project rapid warming to continue [*IPCC*, 2014], threatening to degrade both the areal extent and thickness of permafrost. Many studies predict that ALT will increase in response to this warming

[*Ge et al.*, 2011; *Lawrence et al*., 2012; *Frampton and Destouni*, 2015]. Thawing of permafrost may result in the formation of thermokarst landscapes [*Jorgenson and Osterkamp*, 2005; *Kokelj and Jorgenson*, 2013; *Lara et al*., 2016], change how water is cycled and stored [*Woo*, 1990; *St. Jacques and Sauchyn*, 2009; *Connon et al*., 2015], and alter carbon storage by either enhancing organic matter accumulation or increasing methane emissions [*Turetsky et al*., 2007; *Johnston et al*., 2014; *Helbig et al*., 2016a].

In the zone of continuous permafrost, permafrost thaw occurs predominantly in the vertical direction (*i.e.* increase in ALT), but in warmer environments, where permafrost is discontinuous or sporadic, permafrost thaw can occur both laterally and vertically [*Åkerman and Johannson*, 2008; *McClymont et al., 2013*]. In the southern fringe of discontinuous permafrost where MAAT is close to 0°C, permafrost is typically sporadic or discontinuous (<50% spatial coverage), thin (<10m), relatively warm (>-2°C) and the ALT is relatively large and reported to be increasing [*Brutsaert and Hiyama*, 2012; *Quinton and Baltzer*, 2013]. There has been a documented northward migration of the southern boundary of discontinuous permafrost (*Kwong and Gan*, 1994; *Beilman and Robinson*, 2003) and most climate and permafrost models predict that this trend will continue (*Anisimov and Nelson*, 1996).

ALT is typically measured in the field using a frost probe or simulated numerically using a thermal model. A common misconception is that ALT is equivalent to the depth to permafrost. As a result, ALT field measurements are almost always taken by measuring the distance from the ground surface e to the permafrost table (PT) at the end of summer [*i.e. Brown et al*., 2008; *Mutter and Phillips*, 2012; *Sjöberg et al*., 2012; *Johansson et al., 2013*; *Jean and Payette*, 2014]. Measurement of ALT using this method implicitly assumes that the entire thickness of ground above the PT re-freezes during winter, though re-freeze depths are rarely reported or measured.

In areas of thawing discontinuous permafrost, this entire thickness may not re-freeze, resulting in the formation of a shallow talik (layer of perennially unfrozen ground in permafrost areas) between the overlying active layer and the underlying permafrost. Complete re-freeze can genearlly be assumed in areas of continuous permafrost where low temperatures and abundant ground ice limit thaw (*i.e.* Woo and Young, 2005), however, talik conditions may also exist locally in colder climates where the ground thermal regime and high snow cover can restrict energy loss over winter. For example, using a permafrost thermal hydrology model and simulating future active layer thicknesses, *Atchley et al*. [2016, Fig. 4b] found that taliks may form at Barrow, Alaska under inundated conditions with heavy snowfall.

As the subject of permafrost thaw is gaining interest across disciplines, there is a need to ensure consistency in how the active layer is defined and measured. *Muller* [1947] suggested that the term *suprapermafrost layer* be used to "describe the combined thickness of the active layer and talik", but this term is not currently used in the literature. In areas of discontinuous permafrost, or in areas where it is unknown whether the entire layer above the PT re-freezes over winter, end-of-summer measurements of the depth to the PT are more accurately termed measurements of the suprapermafrost layer thickness (SLT). This term refers to the zone above the PT without implying the presence or absence of a talik. In areas where the suprapermafrost layer freezes entirely, ALT and SLT are synonymous. Changes to ALT and SLT in response to a warming climate are conceptualized in Figure 4-1, which also provides a basic framework that field practitioners can use to identify changes to ALT and SLT based on field measurements.

Most research on taliks has focused on those below aquatic bodies, such as lakes [*Kokelj et al*., 2009; *Arp et al*., 2016; *You et al*., 2017] and wetlands [*O'Donnell et al*., 2011; *Sjöberg et al*., 2016]. Water flowing through taliks can introduce energy for permafrost thaw and can serve to limit downward progression of the freezing front in the winter. This process of energy advection through taliks has been evaluated through numerical simulations [*e.g.* see review by *Kurylyk et al*., 2014]. Some studies have used numerical models to project the development of taliks in terrestrial environments [*Kane et al*., 1991; *Lawrence et al*., 2012; *Atchley et al.*, 2016], however few have documented and studied them in the field [*Jin et al*., 2006; *Mutter and Phillips*, 2012]. The presence of relatively shallow taliks in terrestrial systems is rarely reported.

In this study we present data from a field site in the southern Northwest Territories (NWT), Canada, where measurements indicate that thin taliks (thickness of < 2m) exist below terrestrial systems between the active layer and permafrost. We hypothesize that these shallow taliks facilitate rapid permafrost degradation and, limit perpetual growth of the active layer. The objectives of this chapter are to: 1) demonstrate that the active layer does not always extend to the permafrost table; 2) examine the influence of taliks on thaw rates of the underlying permafrost; 3) document changes to the ground thermal and moisture regime at a rapidly thawing permafrost cored peat plateau.

4.2 STUDY SITE:

Field work was conducted at the Scotty Creek Research Basin (SCRB, 61°18'N, 121°18'W, Fig 4-2a), located about 50 km south of Fort Simpson, NWT in the zone of sporadic-discontinuous permafrost [*Heginbottom*, 2000]. MAAT (1981-2010) at Fort Simpson is -2.8°C, with a mean January temperature of -24.2°C and a mean July temperature of 17.4°C. MAAT has been rising rapidly in the region (2.5°C from 1950 to 2015) with the most pronounced increase in winter (4.5°C from 1950 to 2015) [*Vincent et al*., 2015]. Mean annual precipitation (1981-2010) is 390 mm, with 149 mm (38%) falling in the form of snow. Annual precipitation has remained relatively stable over the past 50 years.

Scotty Creek drains a 152 km^2 subarctic boreal forest watershed in the Taiga Plains ecozone. Field studies were focussed on the wetland-dominated headwaters, where the landscape is dominated by a mosaic of forested peat plateaus and wetlands mainly in the form of treeless collapse scar bogs and channel fens. Permafrost at the SCRB occurs below forested peat plateaus as described by *Zoltai* [1993], while the surrounding wetland terrain of bogs and channel fens is permafrost-free. Permafrost thaw has been rapid in this region [*Robinson and Moore*, 2002]. Permafrost occupied approximately 70% of the Scotty Creek headwater area in 1947, but had decreased to approximately 43% by 2008 [*Quinton et al*., 2011]. As the permafrost below plateaus thaws, the plateau ground surface subsides and becomes inundated by the adjacent wetlands, a process that results in the conversion of forest to wetland, [*Zoltai*, 1993; *Quinton et al*., 2011; *Lara et al*., 2016]. The predominant tree species of the forested plateaus is black spruce (*Picea mariana*) and ground vegetation takes the form of Labrador tea (*Rhododendron groenlandicum*), lichens (*Cladonia* spp.) and mosses (*Sphagnum* spp.).

Peat deposits in this region range from 2 to 8 m in thickness and overlie a silty-clay glacial till [*McClymont et al*., 2013; Connon *et al*., 2015]. As the study site is blanketed by thermally-insulating peat, the permafrost below the plateaus is protected by the large thermal offset between the ground surface and the permafrost [*Robinson and Moore*, 2000; *Smith and Riseborough*, 2002]. Permafrost preserved by peat is a common feature of high boreal and subarctic regions throughout the circumpolar north, especially in the zones of discontinuous and sporadic permafrost [*Brown*, 1970]. A detailed description of Scotty Creek and the hydrological function of each land cover type can be found in *Quinton et al*. [2003].

Measurements of depth to permafrost and plateau width have been taken since 1999 at a 'Study Plateau' (Fig. 4-2d, Transect 5), along with standard meteorological data and subsurface temperature and soil moisture data. The Study Plateau is flanked by a flat bog on one side and a channel fen on the other. It is experiencing rapid lateral and vertical permafrost loss with the current width (<20 m) being less than half of what it was when measurements began in 1999 (~40 m). This Study Plateau allows for a unique investigation of a permafrost peat plateau that is actively transitioning into a wetland.

4.3 METHODS:

4.3.1 SLT and ALT measurements and talik occurrence:

SLT measurements have been taken at the end of summer since 1999 at 1 m intervals along a transect spanning the width of the Study Plateau (Transect 5). In 2011, eight additional transects (widths between 21 and 69 m) were created to determine if the changes occurring at the Study Plateau were representative of the rest of the basin. Measurements at the additional transects were made at 3 m intervals. Each additional transect intersected a wetland/plateau boundary and extended into the interior of a permafrost plateau. Each transect was positioned so it would include the permafrost conditions of both the stable plateau interior, and the 3 to 15 m wide plateau margin where the SLT is significantly greater (Baltzer *et al*., 2014). In 2011, a 40 m x 40 m grid with 5 m measurement intervals (64 points) was established on the interior of a large plateau to measure permafrost thaw without the influence of edge effects (Fig 4-2d).

Each measurement point on the transects and in the grids was marked with a flag and located using a differential GPS system (SR530 RTK, Leica Geosystems Inc., Norcross, USA; system accuracy ±0.02m). Ground surface elevation was also recorded at all points, apart from 15 points in the grid where the GPS signal was lost. SLT measurements were made at all points in late summer (26-29 August) in 2011, 2015 and 2016. Early spring (12-16 May) and late spring (4-6 June) measurements of seasonal thaw (*i.e.* depth to the frost table) at each transect were made in 2011 and 2015. Thaw depth measurements were made by inserting a 1 cm diameter, graduated steel rod (*i.e.* frost probe) into the ground until the depth of refusal [*Nelson and Hinkel*, 2004].

In view of the definition of the active layer, ALT measurements require a measurement of the depth of re-freeze in addition to end of summer thaw depth measurements. Thermistor and soil moisture sensors can provide an approximate depth of re-freeze, but logistical constraints typically prevent a large spatial distribution of these sensors. The depth of re-freeze was measured along the nine transects using either a frost probe or an ice auger in early April 2016.

4.3.2 Frost Probe:

After snow was removed from above each measurement point, a frost probe was inserted into the ground until the depth of refusal. The frost probe was then removed from the ground and re-inserted forcefully and repeatedly in order to penetrate incrementally into the frozen active layer. Each time the frost probe was removed and re-inserted into the ground, the new depth was read and recorded (average increase of < 1 cm per increment). If the depth increased by at least 10 cm in a single increment, it was assumed that the active layer had been fully penetrated and a talik was reached. The last incremental depth value recorded prior to penetrating the talik was assumed to be the depth of re-freeze. The depth to refusal on the next increment was assumed to be the PT. When a talik was not found (*i.e.* ALT = SLT), probing was terminated at the depth equal to SLT as recorded in the previous fall.

4.3.3 Ice Auger:

Measurements taken using the frost probe were verified at 50 randomly selected points using an ice auger. A 15 cm diameter hand-held ice auger (*Rapala*, Oshawa, Canada) was used to drill through the frozen peat until the underlying talik (if present) was reached. The maximum depth of the freezing front was easily identifiable as the interface between frozen and unfrozen peat and can be accurately measured with a ruler. After the talik was reached, a frost probe was used to measure the remaining depth to the PT needed to define the SLT. The water pressure in the talik was always greater than atmospheric, causing the water level in the borehole to rise above the base of the frozen layer. The ice auger method allows for an accurate measure of the depth of re-freeze, however it is destructive, and can be labour-intensive and time consuming. Measurement differences between the two methods were < 5 cm, suggesting that the frost probe method was sufficiently accurate. Measuring the depth of re-freeze using the frost probe is much more practical when many measurement points are required and when destructive sampling is not viable.

4.3.4 Ground Temperature and Soil Moisture:

An array of 11 thermistors (soil temperature) and five water content reflectometers (soil moisture) were installed in a 0.7 m deep soil pit near the centre of the Study Plateau in 2001 (see Hayashi *et al*., 2007 for instrumentation design). The data from these sensors are used in this study to monitor soil temperature and liquid water content. Soil moisture data was used to calculate the total number of days that the soil was frozen at each depth. A sharp decrease in volumetric liquid soil moisture is detected by the water content reflectometers during soil freezing. Typically, water content sensors at depth (0.3, 0.4 and 0.5 m below ground surface) are saturated and have a moisture content of about 0.8 (*i.e.* porosity). Liquid water content in frozen soils at the site is typically in the range of 0.15 to 0.2. To allow for some variation in liquid water content in the frozen state, a frozen condition was assumed during winter when the liquid water content decreased below 0.3.

4.3.5 Ground Heat Flux:

Ground heat flux (Q_G, W m^{-2}) at the Study Plateau was measured using a ground heat flux plate (Campbell Scientific, HFT3) at the soil pit referred to above. The heat flux plate was installed at 0.05 m depth and was calibrated as in *Hayashi et al*. [2007]. Heat storage in the upper 0.05 m of the soil was added to the flux measured by the heat flux plates to obtain a measure of the heat flux at the soil surface [*Mayocchi and Bristow*, 1995]. *Hayashi et al*. [2007] rigorously tested the quality of the ground heat flux plates by comparing these values to those derived from the calorimetric and gradient methods. The authors demonstrated that after applying the calibration, all three estimates yielded similar results.

4.3.6 Radiation and Temperature:

Four component radiometers (Kipp & Zonen, CNR1) were installed 2 m above the ground surface and mounted on a tripod at a thawing peat plateau (Study Plateau), a stable plateau, and in an adjacent permafrost-free bog. Air temperature was recorded at the same locations with a thermistor housed in a Gill radiation shield. All sensors were connected to a data logger (Campbell Scientific, CR10X) which took measurements every minute, and averaged and recorded values either hourly or half-hourly.

4.3.7 Snow Measurements:

Snow water equivalent (SWE) is measured annually at the end of winter along snow courses established in March, 2006. Due to differences in canopy snow interception and snow accumulation in different land cover types, the end of winter snow surveys are conducted independently on plateaus, bogs, fens and lakes. Measurements of snow depth were made at intervals of 1 to 5 m using a ruler and SWE is measured at every fifth depth measurement point using an Eastern Snow Conference snow sampler [*Wright et al.*, 2008] and calibrated scale. SWE values were then averaged for each transect and land cover type for each year.

4.4 RESULTS AND DISCUSSION:

4.4.1 Talik Occurrence:

Our measurements indicate that points with greater SLT do not necessarily have greater ALT, demonstrating the need to take measurements of the depth of re-freeze (Fig. 4-3). These data also demonstrate that the likelihood of a talik at a given site can be evaluated from end of summer SLT measurements at this monitoring site. Specifically, when SLT is < 60 cm, the ground completely re-freezes during winter, preventing talik formation. The maximum measured ALT among the 135 measurements was 74 cm. Considering the < 5 cm error associated with the frost probe method, a talik can be reasonably assumed when the SLT is greater than approximately 80 cm. For the 60 to 80 cm depth range, a talik may or may not be present. These findings indicate that there is a threshold for maximum ALT, preventing the perpetual growth of the active layer over time. To our knowledge, the concept of a threshold for maximum ALT has not yet been discussed in the literature. The data and locations of vertical boundaries plotted on Figure 4-3 represent a single annual period at Scotty Creek, and may vary year to year with variations in air temperatures, timing and magnitude of snowfall and changes in soil moisture. Although the methods used here are applicable to other permafrost regions, site specific measurements are required to evaluate local ALT values.

Similar to ecological space-for-time substitutions [*Blois et al*., 2012], Figure 4-3 suggests how SLT and ALT might change at a specific point or site over a period of decades while the underlying permafrost thaws. Initially, active layer thickening is expected, as the ALT increases from 40 to 60 cm. As the SLT increases to between 60 and 80 cm, ALT may be highly variable inter-annually and may be governed by either maximum depth of thaw or freezing, depending on meteorological conditions. Eventually, when the ~80 cm ALT threshold is surpassed, ALT is controlled entirely by the maximum depth of freezing.

4.4.1.1 Active Layer Thinning:

In response to higher winter air temperatures in northern Canada [*Vincent et al*., 2015] and/or increases in snowfall [*Prowse and Furgal*, 2009], freeze-thaw models predict a thinning of the layer which re-freezes annually [*Fraunfield et al*., 2004; *Lawrence et al*., 2012]. Assuming saturated conditions, the heat capacity of ground with a talik is greater than ground without a talik. The talik will retain more energy and experience a temperature increase (in the form of sensible heat). As this energy must be removed before re-freeze can commence, re-freeze in areas with taliks may lag those without taliks. This results in a delayed onset of the zero curtain period (period of time in which the ground remains isothermal at the freezing point due to phase change). This is compounded when early season snow accumulation insulates the ground from cold atmospheric temperatures and restricts energy loss. Furthermore, higher air temperatures in winter will also decrease the thermal gradient between the atmosphere and ground surface. This combination may limit the downward penetration of the freezing front, and in turn, decrease ALT. Therefore, in areas where ALT is governed by the maximum annual depth of the freezing front, ALT should decrease in response to higher temperatures. This is supported by the data presented in Figure 4-3 which shows that ALT decreases slightly as the SLT increases. For example, ALT of < 40 cm are observed only where SLT > 1 m. To the knowledge of the authors, the concept of a thinning active layer in response to a warming climate has not yet been discussed in the literature.

4.4.1.2 Talik Classification

Shur et al. [2005] discussed the importance of classification when describing the vertical profile of a permafrost system, and demonstrated the need to include the transition zone (top layer of permafrost that undergoes freeze-thaw on longer timescales) in conceptual models. Expanding on this idea, we recommend including the suprapermafrost layer and taliks in such models. In the absence of a talik, permafrost aggrades, degrades or remains stable by the balance between energy gained during summer when the thermal gradient is directed downward, and energy lost during winter when the thermal gradient is directed upward. However, once a talik has formed, the permafrost body continually gains energy from the downward directed thermal gradient of the relatively warmer talik during winter and throughout the year. The presence of a talik also introduces an upward-directed flow of energy to the active layer. If active layer measurements are used to validate numerical models [see *Brown et al*., 2008, pp 168], it is necessary to differentiate between the active layer and suprapermafrost layer (when applicable) to accurately represent subsurface thermodynamics.

4.4.2 Effects of taliks on permafrost thaw

These observations indicate that once a talik is formed, thawing of the underlying permafrost accelerates. Increases in SLT are substantially greater at points that have a talik than at those without (Fig 4-4). The median SLT increase over five years (2011 to 2016) was 37 cm at points with a talik, but only 8 cm at points without a talik. Measured increases in SLT over the one year (2015 to 2016) period indicate a median SLT increase of 7 cm at points with a talik, but only a 1 cm increase at points without a talik.

A frequency distribution (Fig. 4-5) compares thaw depths at all transects for early spring, late spring and late summer between 2011 and 2015. Late summer thaw depth measurements for the grid are also included for the same two years. Q_G during the summer period (01 May to 31 August) of 2011 and 2015 was similar (183.4 MJ m^{-2} and 179.3 MJ m^{-2} respectively), as was the location of the frost table relative to the ground surface in early spring (Fig. 4-5). If it is assumed that the frost table defines the top of the saturated, frozen layer, then ground ice content is similar between the two years as well. This would suggest that average end of season thaw depth should also be similar between the two years in a stable system. The distribution of early and late spring thaw depths are similar, however the distribution of late summer thaw depths vary significantly. In 2011, only 20% of points on the transects had an SLT > 80 cm, compared to almost half of the points (48%) in 2015. Assuming points with an SLT > 80 cm have a talik (Fig. 4-3), there is a drastic increase in the number of points with a talik over the course of four years. This trend is also observed when considering data from the grid. In 2011, 8% of the grid had an SLT > 80 cm, whereas by 2015 this value increased to 33%. The proportion of points in the grid with taliks is likely smaller than the proportion of points on the transects with taliks because the transects traverse the edges of plateaus (see Fig. 4-2c), which are more likely to have increased thaw depths [*Baltzer et al*., 2014]. The grid is located entirely on the interior of a plateau and is devoid of these edge effects.

4.4.2.1 A "Tipping Point" type change:

We hypothesize that the dramatic changes occurred to ground thermal regimes at the study site between 2011 and 2015, and that these constituted a type of tipping point that generated large increases in the rate of talik development and expansion. The net annual ground heat flux (Fig. 4-6) in 2012 (108.9 MJ m^{-2}) was the largest since measurements began in 2003, and the summer contribution (192.4 MJ m^{-2}) was among the highest as well (average: 170.6 MJ m^{-2}; SD: 18.6 MJ m^{-2}). The following winter, end of season SWE on the plateaus was 151 mm, much higher than the average (127 mm) of the 13 year record at Scotty Creek. Together, the abnormally high summer energy input combined with the insulating effects of increased snow cover inhibited penetration of the freezing front during the winter of 2012/2013. This is likely to have initiated the talik development documented in this study. The soil moisture data (Fig. 4-7), also shows that 2012 was the first year of incomplete re-freeze at the 40 and 50 cm depths. In three of the four years following 2012, re-freeze does not occur at 50 cm depth. This suggests that one abnormally warm thaw season, followed by a winter with shallow re-freeze depths can cause widespread talik development leading to rapid permafrost thaw. The recurring net annual positive Q_G contributions indicate that permafrost in this region is in disequilibrium with the current climate and is especially susceptible to degradation during warm years.

Chasmer and Hopkinson [2016] postulate that the strong El Nino/Southern Oscillation (ENSO) event of 1997/1998 may also have been a tipping point for accelerated permafrost thaw in areas of discontinuous permafrost. They show that permafrost coverage and runoff ratios change significantly after this event, however they do not identify a mechanism for this expedited thaw. We propose that similarly to the warm summer of 2012, the ENSO event may have initiated talik development at some locations within the basin. Unfortunately field monitoring at the SCRB did not begin until 1999, so field based analysis of the impact of the ENSO event on thaw depth is not available. It has been well documented that permafrost exerts control on water cycling and storage [*Carey and Woo*, 1999; *Wright et al*., 2009], and that permafrost thaw may change these patterns [*St. Jacques and Sauchyn*, 2009; *Connon et al*., 2014, 2015; *Walvoord and Kurylyk*, 2016). As shown in this study, talik formation results in more rapid permafrost thaw, and may in turn make the basin more effective at conveying runoff and thereby increase runoff ratios.

4.4.3 Case study of a rapidly thawing permafrost plateau:

Ongoing field measurements since 1999 have documented vertical and lateral retreat of permafrost at a long-term Study Plateau in the SCRB (Fig. 4-8). Thaw rates at eight additional transects indicate that this rapid permafrost loss is not consistent across the sampled portion of the basin (Table 4-1). The Study Plateau transect is underlain by a talik that provides a subsurface connection between the bog and the channel fen (as shown in cross sections Fig. 4-2c). The additional eight transects do not cross any other taliks connecting wetlands, however additional probing in August 2016 found other locations in the basin with a fully connected talik between two wetlands. All features were found in areas where the plateau width between two wetlands is < 30 m. The data presented in this study show that the study transect has undergone nearly full conversion from a raised peat plateau to a wetland. Initial SLT measurements at the Study Plateau (1999-2004) suggest that taliks were not present on the plateau when it was first instrumented (Fig. 4-6), and subsequent measurements illustrate the growth and expansion of taliks. This plateau offers a unique opportunity to document changes to subsurface soil properties as the permafrost thaws over time.

4.4.3.1 Ground Heat Flux Changes:

Net annual Q_G was positive for 8 of the 11 years in the study (Fig. 4-6), providing unstable conditions for the underlying permafrost. 2004 had the lowest net ground heat flux (-128.6 MJ m^{-2}) and was the year with the lowest recorded average SLT. Other years with a small negative annual ground heat flux (2007 and 2011) did not respond with decreases in SLT, nor did they experience the appreciable increases in SLT observed in other years. During these years there was enough energy to thaw the active layer, but not enough to thaw the underlying permafrost. The year with the largest Q_G (2012) was the first year that the water content reflectometers indicated incomplete freezing of the soil. In 2015, a relatively small annual Q_G resulted in the second largest increase in SLT. This suggests that downward conduction of energy may not be the only mechanism contributing to soil thaw. *Quinton and Baltzer* [2013, Fig 6c] plotted thaw curves for each year from 2002 to 2010 at each of the 11 thermistor depths. These curves were relatively linear (*i.e.* unattenuated) throughout the soil profile, despite a decreasing thermal gradient as the thaw front extended further below the ground surface. The reason for the lack of attenuation is unclear, however advection of energy through the connecting talik may provide an additional energy source to expedite permafrost thaw. Further investigation is required to quantify the contribution of advection to permafrost degradation.

There is a demonstrated link between moisture levels and thaw depth [*Hayashi et al*., 2007; *Wright et al*., 2009, Figs. 6 and 7]. In 2006 the point with the highest elevation on the plateau along the Study Plateau transect was 0.9 m higher than the adjacent wetlands [*Wright et al*., 2009]. In 2015, the highest point was only 0.42 m higher than the wetland surface, indicating subsidence of the plateau in response to vertical permafrost thaw. As the ground surface subsides, it becomes closer to the elevation of the water table in the wetlands, and the thermally insulating unsaturated zone thins. This gradual subsidence is consistent with increases in near surface soil moisture conditions, which increases seasonal thaw through an increase in thermal conductivity. When the elevation of the permafrost table at all points along the transect drops below the elevation of the water table in both wetlands, water movement through the plateau is initiated in the direction of the hydraulic gradient. The combination of subsidence, increased moisture levels, talik development and cross-plateau flow enhances permafrost thaw on the plateau, and will result in the eventual conversion to a wetland.

4.4.3.2 Radiation and Soil Moisture Changes:

Ongoing permafrost thaw and talik development at the Study Plateau has resulted in marked changes to subsurface thermal and moisture conditions. Annual maximum subsurface temperatures have been increasing since 2001, with the most significant changes occurring at the three deepest thermistor depths (50, 60 and 70 cm below the ground surface, Fig. 4-9). For example, the maximum temperature at 70 cm depth has increased by 6.7°C over the last 15 years. Soil temperatures at the three depths shown in Figure 4-9 do not drop below the zero curtain for three of the last four years, indicating the presence of a talik in which the soil does not completely freeze. Analysis of soil moisture data shows a decline in the number of days the soil is frozen each year at all depths (10, 20, 30, 40 and 50 cm below the ground surface, Fig. 4-7). The changes documented here may be caused by greater energy input through increased net radiation, less energy loss through increased snowpack, or greater thermal storage in a thickening talik. It has been suggested that in discontinuous permafrost terrains, shading provided by the tree canopy reduces incoming shortwave radiation (K↓) enough to maintain permafrost, where otherwise it would degrade [*Shur and Jorgenson*, 2007]. As the permafrost core of a plateau thaws and retreats laterally, biomass at the plateau edges is reduced [*Chasmer et al.*, 2011], thereby increasing K↓, and the amount of energy incident at the ground surface. This effect is pronounced at high latitudes, where the low sun angle increases the shading influence of trees at plateau edges.

To investigate this, total K↓ on the plateau was compared to K↓ in an adjacent treeless bog on cloud-free days in 2005, 2007 and 2015. It was found that K↓ on the plateau relative to the bog had increased significantly between the two periods (Fig. 4-10), aiding in the thaw of the plateau. In 2005, the Study Plateau received, on average, 80% of the total K↓ that the bog received, as opposed to 87% in 2015. Decreases in late summer plateau to bog K↓ ratios are reflective of lower sun angles at this time of year. K↓ was also compared between a stable plateau and the bog between 2007 (the first year measurements were made) and 2015 and no significant differences in K↓ were found, providing evidence that the changes in K↓ ratios at the Study Plateau are a response to thawing permafrost. Increased K↓ ratios at the Study Plateau suggest a reduction in LAI, allowing for more shortwave radiation to penetrate through the canopy to the ground surface (see half hour resolution in Fig 4-10). It can be reasoned that this canopy thinning may also reduce the interception of snow in the tree canopy, allowing for more snow accumulation on the ground. Increasing snow cover (during early winter) will reduce the amount of over-winter energy loss from the ground and may explain the warmer soil temperatures and talik development.

4.4.3.3 Talik Distribution:

Baltzer et al. [2014] demonstrate that thaw rates around the periphery of plateaus are much higher than on the interior, and conclude that the fragmentation of plateaus will cause faster thaw at the landscape scale. We hypothesize that similar processes occur along the periphery of taliks in the plateau interior and that plateaus may begin to fragment internally, which is consistent with the results of geophysical imaging indicating the presence of deep (ca. 2m) permafrost table along the edge of the Study Plateau [*McClymont et al*., 2013, Fig. 4b]. Within a plateau, taliks can appear as either isolated or connected features (see insets in Fig. 4-2c). We suggest that taliks initially occur on plateaus in areas with a sparse tree canopy, increasing the net radiation incident to the forest floor and enhancing summer thaw depth. If the ground loses insufficient energy over winter to completely re-freeze to the PT, then an isolated talik is formed. This process is analogous to the formation of collapse scar bogs as described by

Robinson and Moore [2000]. Once multiple isolated taliks form on a plateau, their growth and coalescence enables subsurface connections among them and may facilitate perennial groundwater flow. Such a talik will eventually form a subsurface hydrological connection with adjacent wetlands where vertical thaw rates may be enhanced by the lateral transfer of energy via advection [*i.e. Rowland et al*., 2011; *Sjöberg et al*., 2016]. The cross section of the study transect in 2006 (Fig. 4-8) shows a depression in the centre, possibly a remnant of an isolated talik as demonstrated by geophysical imaging [*McClymont et al*., 2013]. We predict that the processes observed and documented on the Study Plateau are indicative of what may happen in other areas of the basin given further warming and thawing of permafrost in this region.

4.5 CONCLUSION:

Active layer measurements have traditionally been made at the end of summer and are assumed to be analogous with the depth to the permafrost table. However, this assumption is falsified by the present study which demonstrates the presence of a talik in the suprapermafrost layer. In discontinuous and sporadic permafrost environments, the vertical profile of a permafrost system should be recognized as potentially containing three layers: the active layer, in which soil pore water changes phase at least twice annually, the talik, in which pore water is perennially unfrozen, and the permafrost, where temperatures are ≤ 0°C for two or more years. This study defines a maximum threshold for active layer thickness, beyond which, the ground is incapable of losing enough energy to completely re-freeze the following winter. It is reasonable to conclude that the thaw and eventual disappearance of permafrost from Scotty Creek will be accompanied by a thinning and eventual disappearance of the overlying active layer (Fig. 4-1b). This thinning will be the result of shallower frost penetration during warmer winters, combined with greater heat storage in taliks as they thicken. This study demonstrates that wide-spread talik development can be induced by a summer with greater than average thaw depth followed by a winter with high snowfall. Once developed, the permafrost underneath taliks thaws much faster than in areas without taliks. This suggests that warm years may generate an abrupt and non-linear response (*i.e.* tipping point) facilitating the simultaneous rapid development of taliks and thaw of permafrost.

4.6 FIGURES

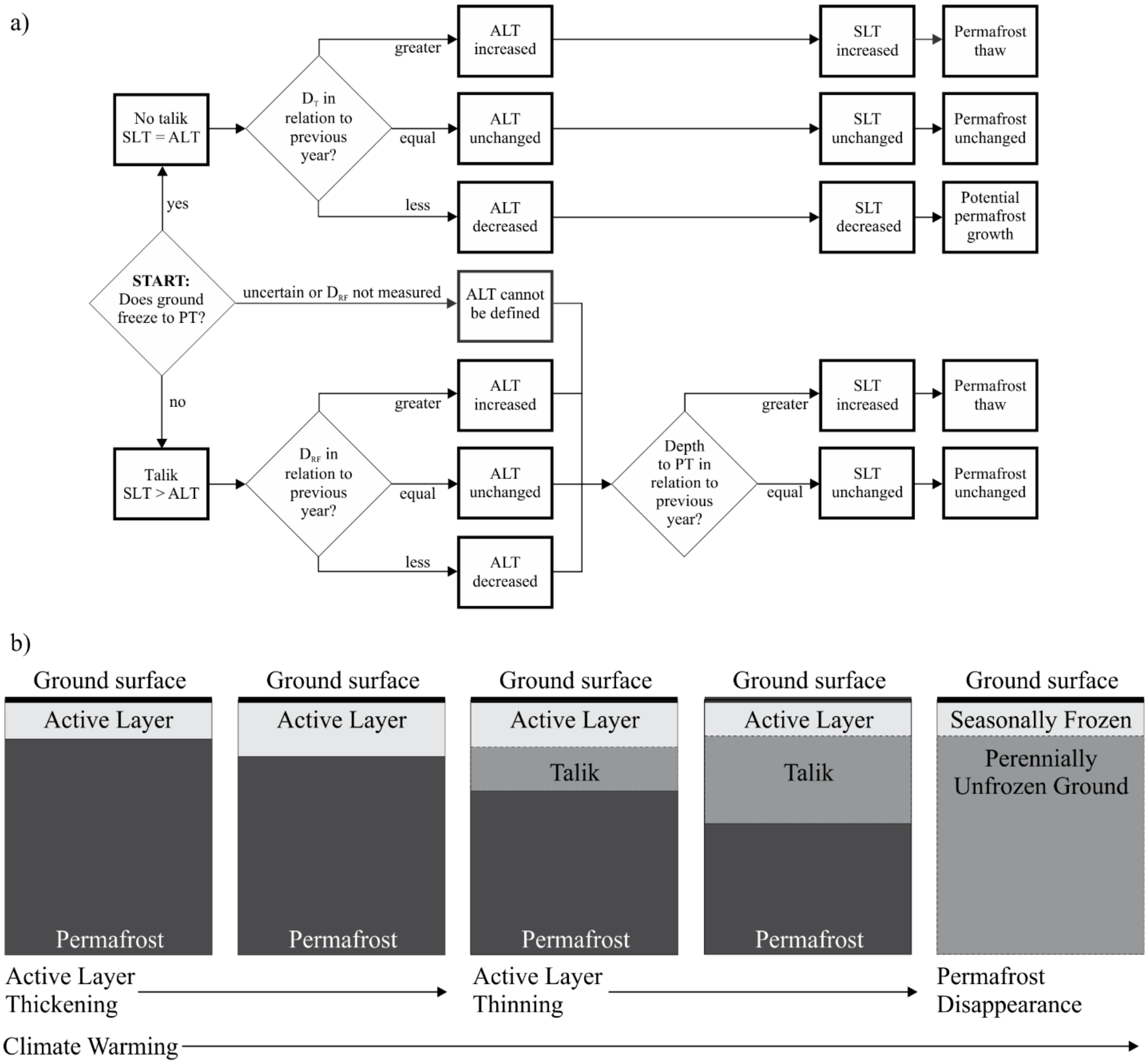


Figure 4-1: a) Conceptual diagram of suprapermafrost layer thickness (SLT) and active layer thickness (ALT) changes based on field measurements and the associated impact on permafrost. D_{RF}: depth of re-freeze, D_T: depth of thaw; ALT: active layer thickness, SLT: suprapermafrost layer thickness, PT: permafrost table; b) Changes in ALT in response to climate warming. ALT reaches a maximum prior to talik development. SLT is the combined thickness of the active layer and talik.

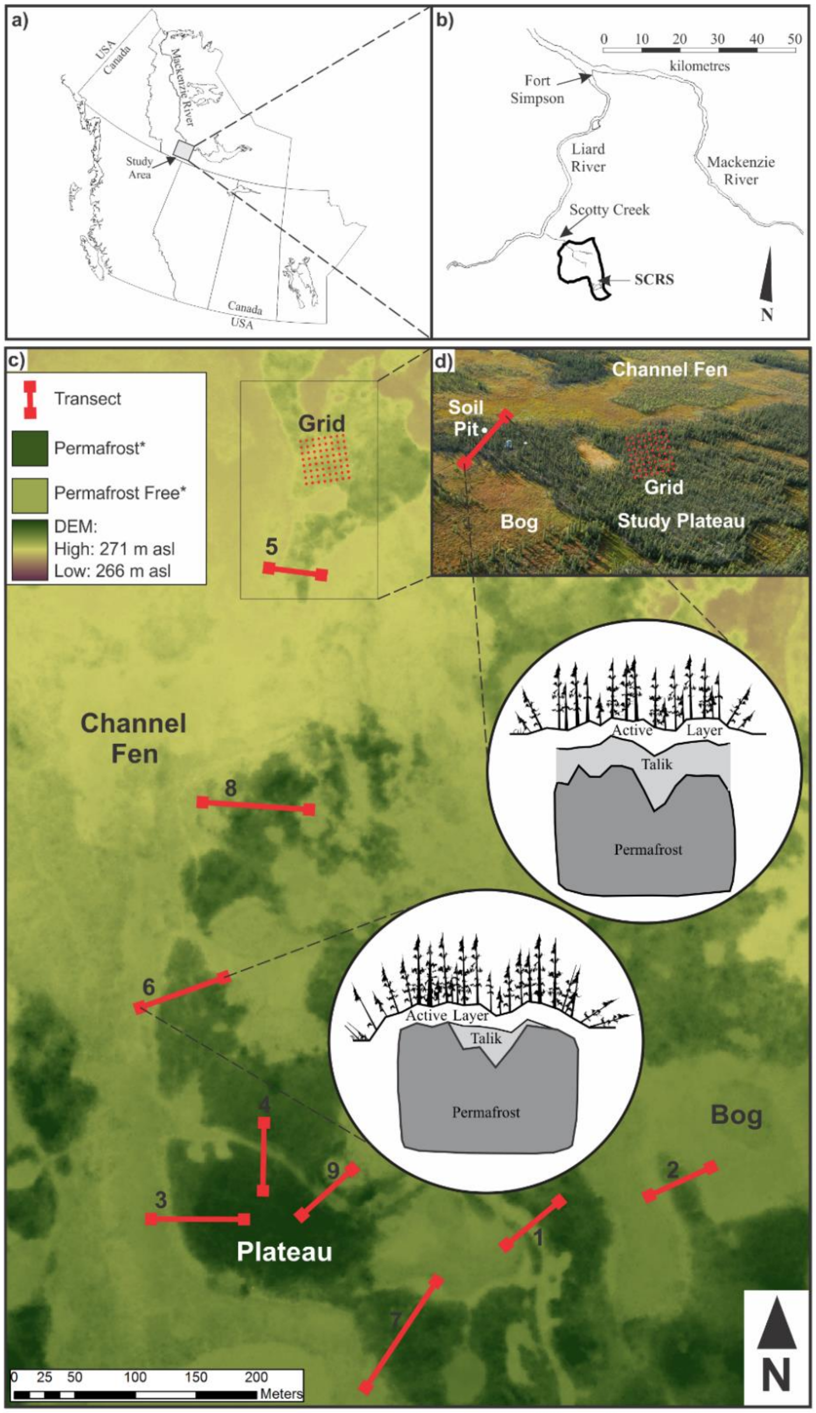


Figure 4-2: a) Map of the study area within Canada; b) Inset of the Scotty Creek region. Bolded basin boundaries represent the area of the basin gauged by the Water Survey of Canada; c) Study site, including nine transects and one grid overlain on a digital elevation model (DEM) of the basin. *Colours for permafrost and permafrost-free terrains are approximate, as colour is based on the DEM and elevation decreases northward. Bubble insets show cross sections at two transects. Ground surface, active layer, talik and permafrost boundaries are all to scale and were measured in the field. The base of the permafrost is for illustrative purposes only and is not to scale. Permafrost thicknesses at the SCRB range from 5 to 13 m [McClymont *et al.*, 2013]; and d) Inset of Study Plateau from aerial photograph (2006).

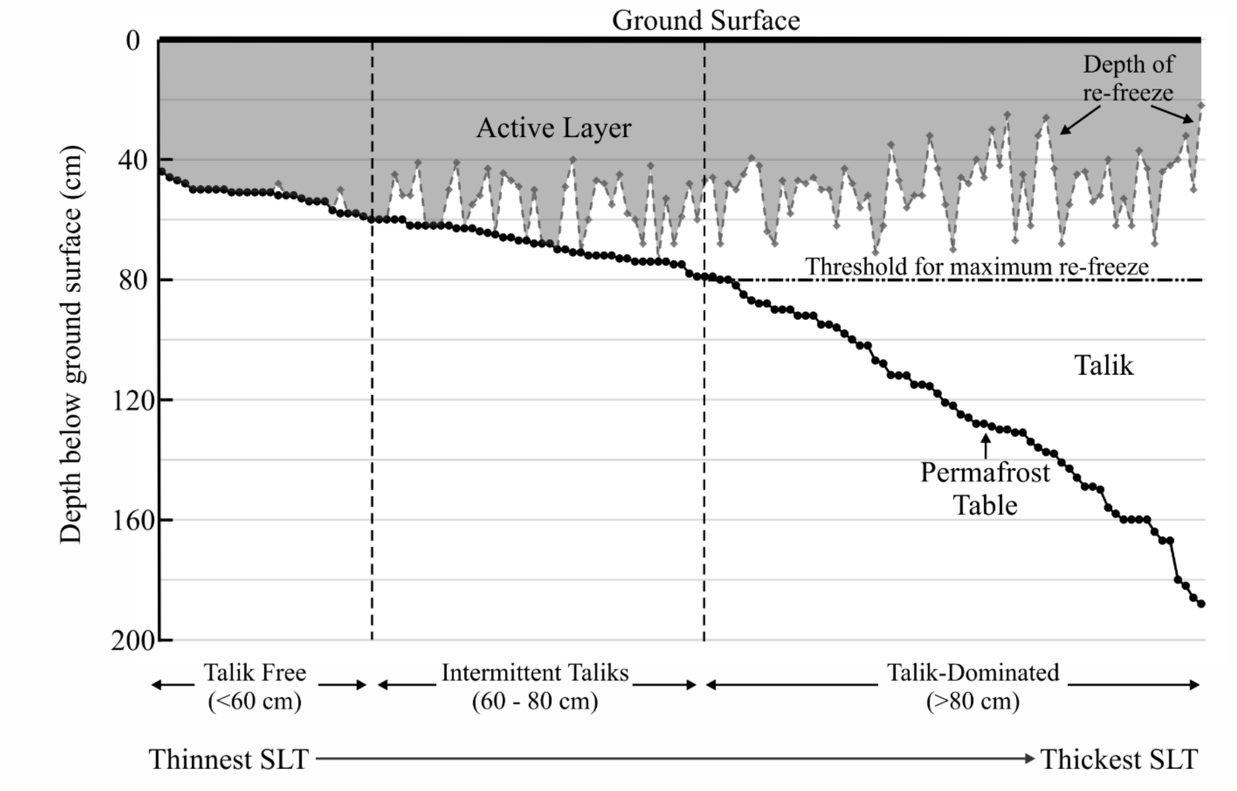


Figure 4-3: SLT measurements ranked from thinnest (left) to thickest (right) and associated thaw depth measurements. Active layer re-freeze depth was measured in April 2016 and thaw depth was measured in August 2015.

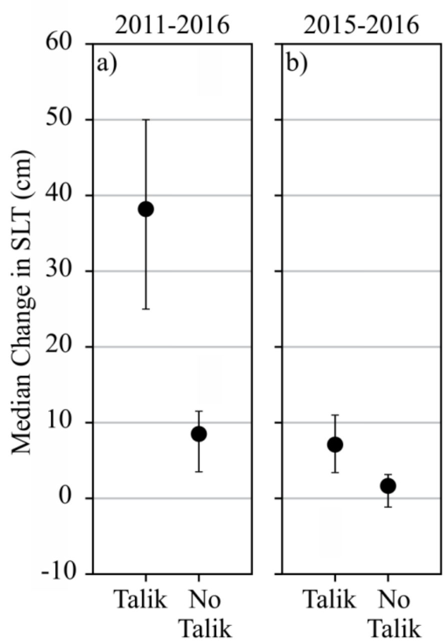


Figure 4-4: Change in median SLT over: a) 5 years (2011-2016); and b) 1 year (2015-2016). Error bars indicate interquartile range.

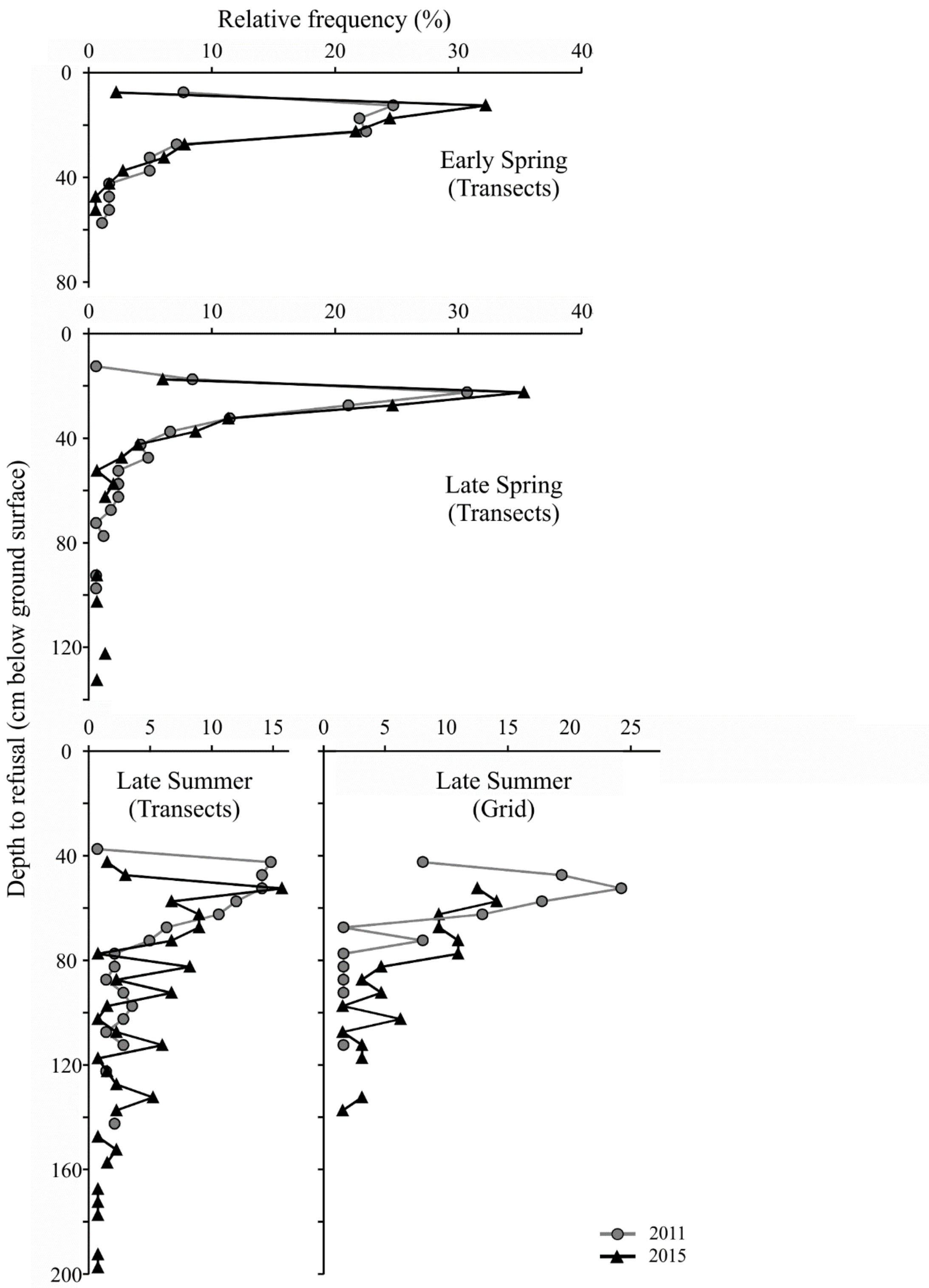
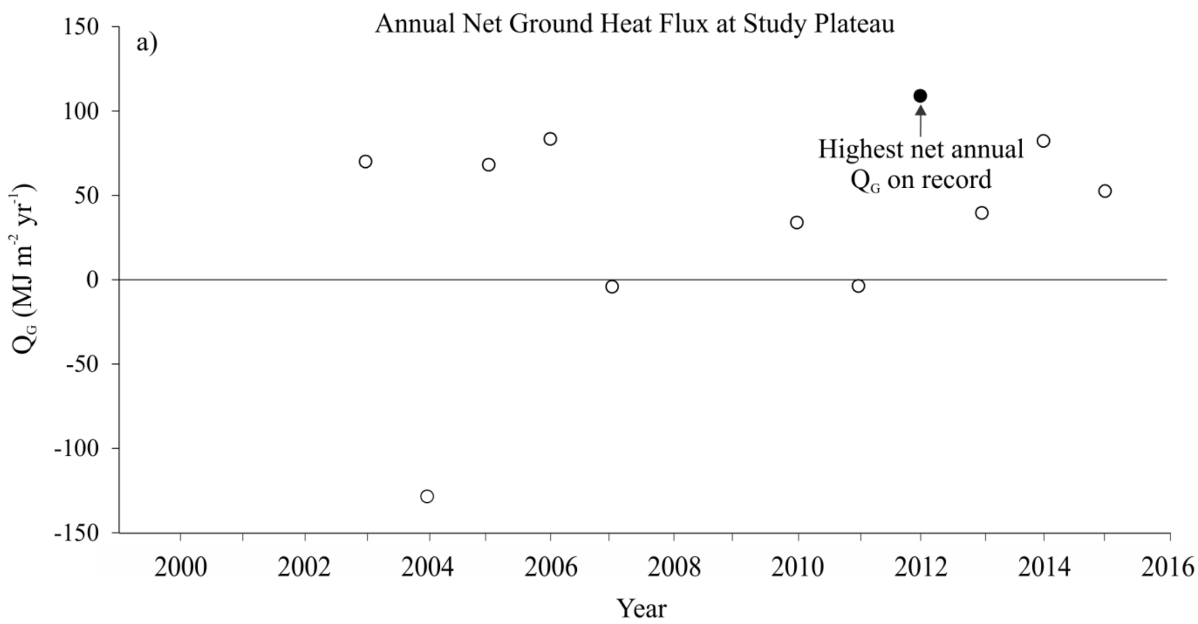


Figure 4-5: Frequency distribution of thaw depths at all transects for early spring (12-16 May), late spring (04-06 June) and late summer (26-29 August) of 2011 and 2015. Late summer thaw depths for the grid are also shown. Of note is the increase in late summer thaw depths of greater than 80 cm for 2015, suggesting increased talik development.



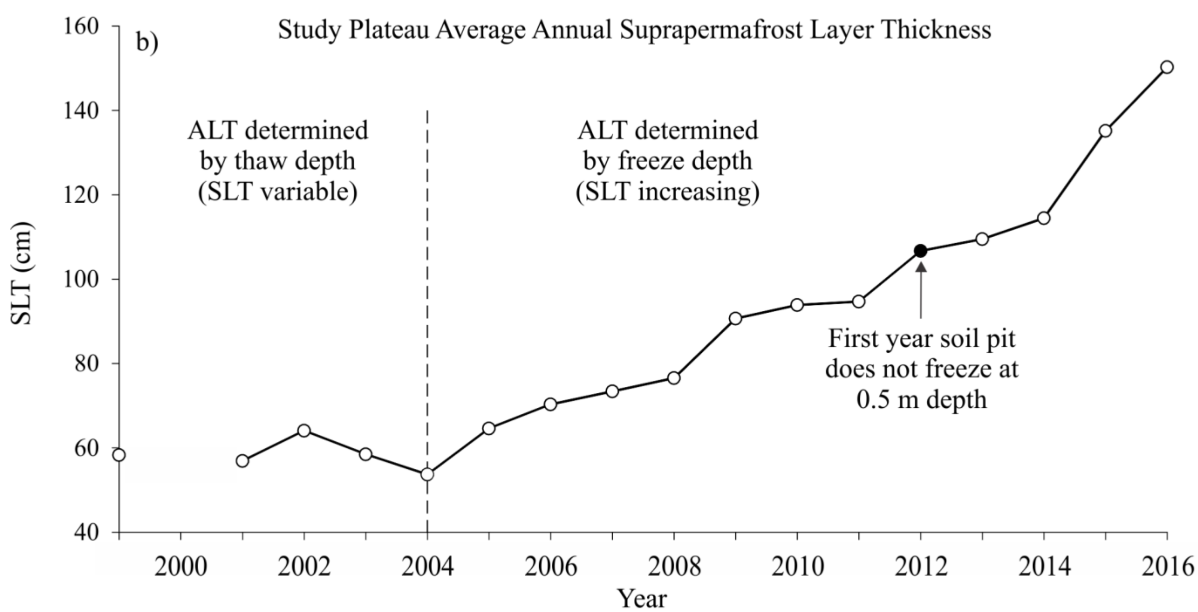


Figure 4-6: a) Net annual ground heat flux measured by a ground heat flux plate at the Study Plateau since 2003. Instrument malfunction prevented Q_G data collection in 2009; b) SLT measured at the end of summer at the Study Plateau.

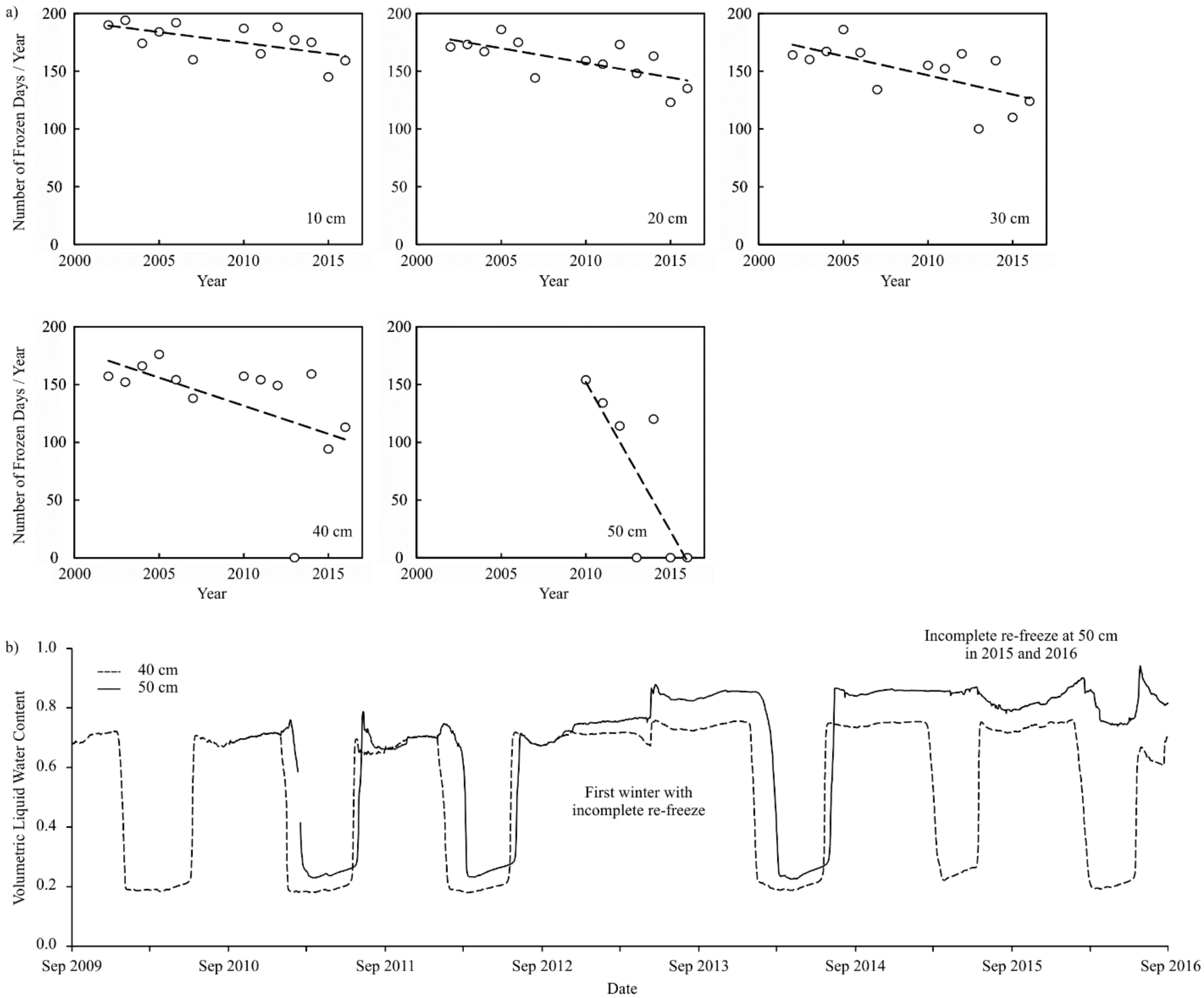


Figure 4-7: a) The annual number of frozen days at 10, 20, 30, 40 and 50 cm depth at the Study Plateau as indicated by the water content reflectometers. A frozen day is assumed when volumetric liquid water content (VWC) dropped below 0.3; b) VWC at 40 and 50 cm depth at the Study Plateau from September, 2009 to 2016. Soil freezing can be observed as periods when VWC drops to values at ~0.2. During years when VWC does not drop, the soil remains saturated with liquid water year round and a talik is assumed.

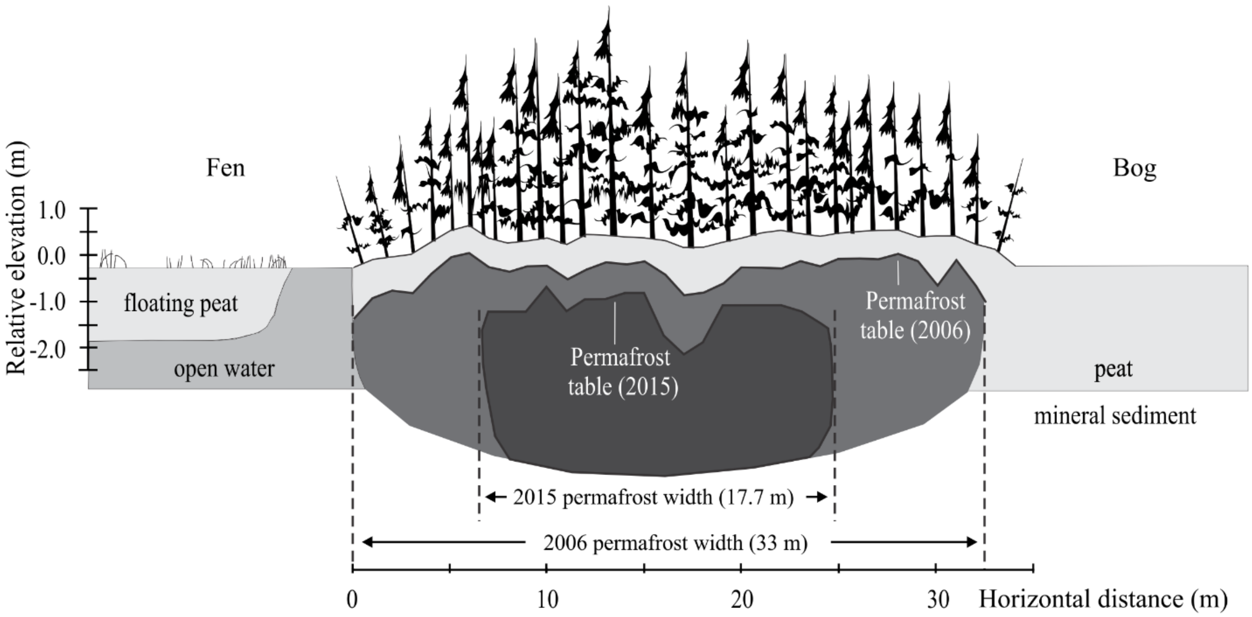


Figure 4-8: A cross section of the Study Plateau depicting lateral and vertical loss of permafrost between 2006 and 2015.

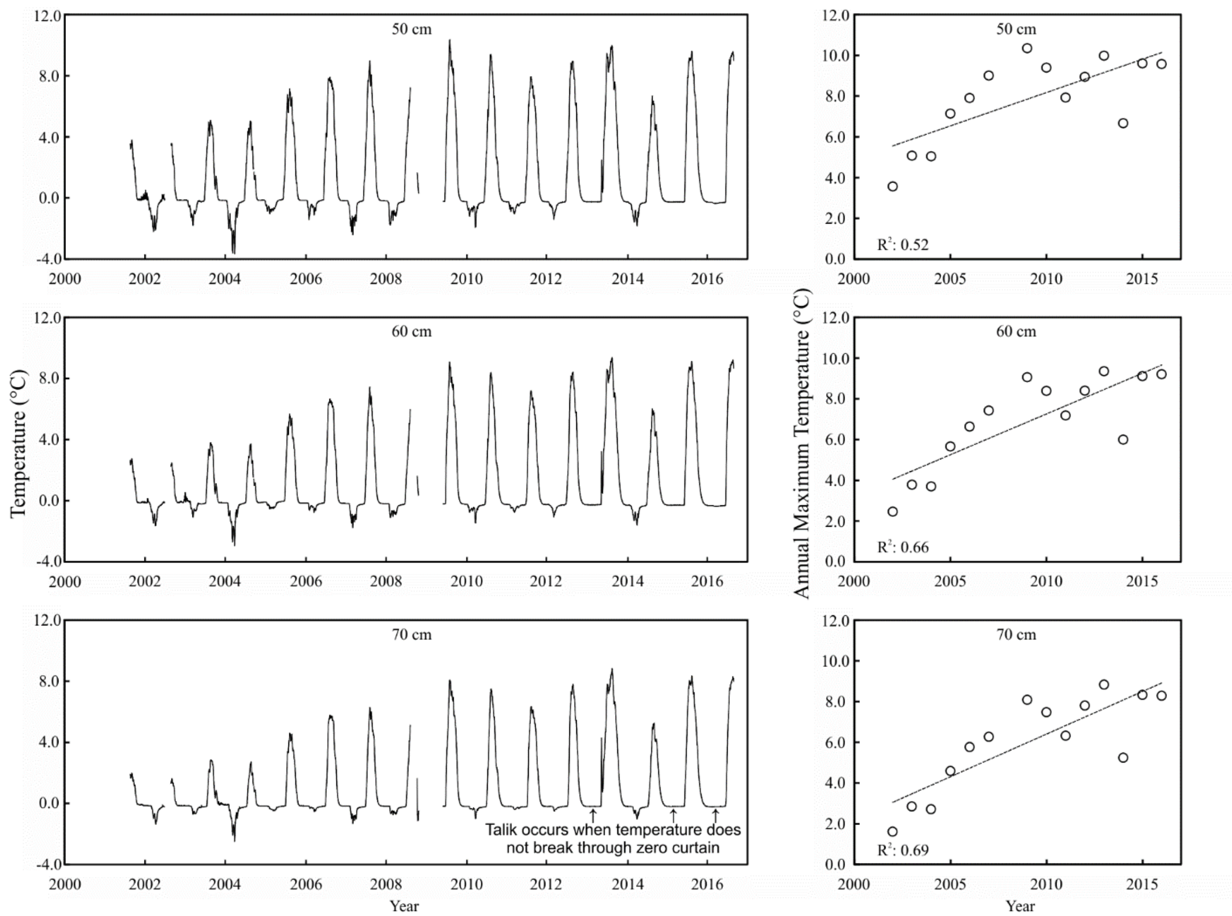


Figure 4-9: Soil temperatures at 50, 60 and 70 cm below the ground surface at the Study Plateau. Annual maximum temperatures have been increasing steadily since 2002.

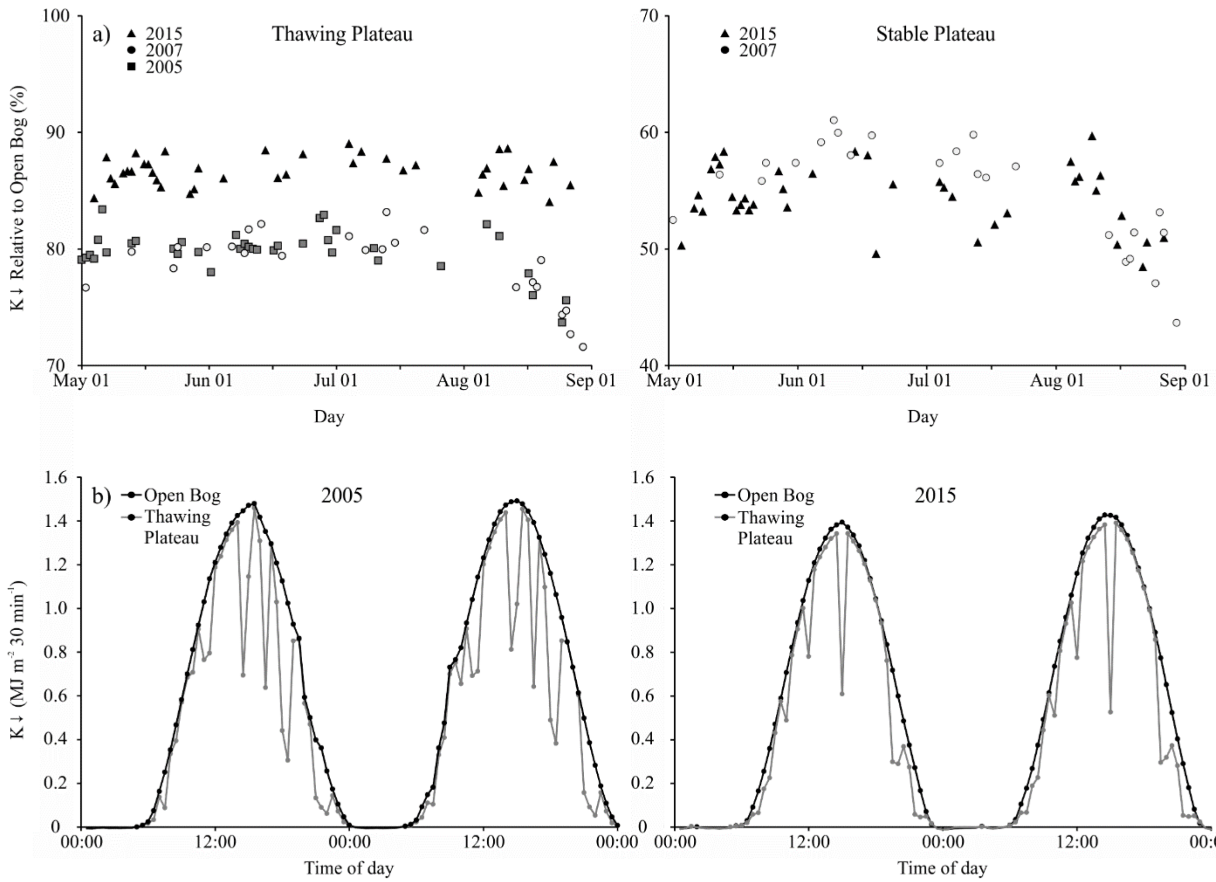


Figure 4-10: a) Comparison of incoming shortwave radiation (K↓) between a thawing plateau (Study Plateau) and a treeless bog and a stable plateau and a treeless bog. Data shown is for cloud-free days in 2005, 2007 and 2015. Note that the stable plateau radiometer was installed in 2007; b) Half hourly K↓ for back-to-back cloud-free days in 2005 (June 8-9) and 2015 (May 18-19) for a thawing plateau and a treeless bog. Periods when K↓ on the plateau decrease relative to the bog are likely periods when the sensor is shaded.

4.7 TABLES

Table 4-1: Average SLT at each transect and grid in 2011, 2015 and 2016 and the % cover with a talik. Taliks were assumed when SLT > 80 cm.

	Average SLT (cm)				% With Talik		
Location	2011	2015	2016	Change	2011	2015	2016
Transect 1	53.1	71.4	73.2	20.1	8%	25%	25%
Transect 2	77.5	85.1	90.6	13.0	25%	43%	43%
Transect 3	57.1	66.5	70.7	13.6	0%	24%	25%
Transect 4	65.5	84.1	83.8	18.3	13%	36%	36%
Transect 5	90.0	135.2	149.7	59.7	73%	100%	100%
Transect 6	58.2	69.6	72.0	13.8	13%	33%	33%
Transect 7	48.5	67.3	70.8	22.2	0%	13%	22%
Transect 8	57.3	82.4	87.2	29.9	10%	45%	58%
Transect 9	63.7	84.6	95.3	31.6	14%	68%	76%
Grid	57.4	76.7	74.7	17.4	6%	33%	38%

CHAPTER 5:
Conclusions and Future Research

5.1 SUMMARY AND CONCLUSIONS:

Field research conducted at Scotty Creek has shown that each of the major land cover types in the basin (*i.e.* permafrost-cored peat plateaus, channel fens and flat bogs) each have a specific hydrological function (Hayashi *et al*., 2004; Quinton *et al*., 2003). Permafrost thaw in the region is rapid (Quinton *et al*., 2011), and is changing the relative proportion of land cover types. The research conducted here documents how permafrost thaw will affect the flux and storage of water in wetland dominated basins in the zone of discontinuous permafrost. Few studies have developed a process-based understanding of how runoff generation processes may change at the local and regional scales. Knowledge of these processes will improve our ability to represent these systems in numerical models and advance our predictive capacity. The processes and data documented here will provide valuable information for parameterization of physically based models.

The hydrologic function of collapse scar bogs is shown to be different than what has been reported previously in the literature. Typically, these features have been regarded as storage features (Quinton *et al*., 2003), as they appear as depressional features on the landscape and are impounded by impermeable permafrost. Collapse scar bogs at Scotty Creek are exhibiting radial growth patterns, coalescing with other wetlands and forming direct surface and subsurface hydrologic connections with the basin drainage network (channel fens). A major theme of this research addresses how changing hydrologic connectivity is increasing the basin runoff contributing area.

Wright *et al*., (2008) elucidate the mechanisms associated with generating runoff from a single plateau; the research here presents a framework for how water is shed from, and routed through a peat plateau – bog complex. A framework of primary and secondary flowpaths is presented, which allows two possible routes for water to reach the drainage network. Primary runoff occurs on the hillslopes that feed directly into a channel fen. Conceptually speaking, this runoff occurs on the periphery of peat plateau – bog complexes. Primary runoff is relatively constant and is dependent on the hydraulic gradient between the plateau and fen and the hydraulic conductivity of the saturated flow zone. As described in Wright *et al*., (2008), primary runoff can be calculated using the Dupuit-Forchheimer approximation.

Bog capture is a process whereby permafrost thaw directly increases the runoff contributing area by removing a permafrost dam that previously prevented drainage. This process may be viewed as analogous to the draining of thermokarst lakes initiated by retrogressive thaw slumps (Segal *et al*., 2016; Marsh and Neumann, 2003). Increases in runoff contributing area are shown to produce increases to the basin runoff ratios, indicating that the basin becomes more effective at transmitting precipitation inputs. Increases to runoff ratios in the lower Liard River valley are significantly correlated to the per cent cover of bog in the basin. Therefore, if a basin has a higher areal coverage of bogs, it can be expected that thawing permafrost will initially produce higher runoff.

Secondary runoff occurs internally on a peat plateau – bog complex and is routed through a bog cascade before reaching the drainage network. The routing occurs through drainage channels that cut through intervening plateaus and is highly dependent on storage conditions. Hydrological routing through a bog cascade can be described using the element threshold concept (Spence and Woo, 2003), where the storage capacity of each bog must be satisfied before that bog is capable of transmitting water further down the cascade. Secondary runoff is therefore inherently tied to the moisture conditions of the system. When moisture levels are high and secondary flowpaths are active, the contributing area in the basin can increase substantially. During these periods, such as the spring freshet, or in response to rain events, the entire bog cascade is capable of transmitting water to the drainage network. Large bogs with low plateau:bog ratios may act as 'gatekeepers' (Phillips *et al*., 2006) and prevent through-flow from upstream bogs during periods of low moisture input. As permafrost thaw continues, it can be expected that secondary flowpaths will become more predominant at the expense of isolated areas. Isolated areas occur where collapse scar ombrotrophic bogs do not have a surface or subsurface connection with another wetland and losses are strictly through evapotranspiration.

Together, chapters two and three combine to show how thawing permafrost can increase the proportion of the basin that is capable of producing runoff. Rates of permafrost thaw between 1977 and 2010 are quantified in chapter two, however this does not provide an estimate as to the rate of permafrost thaw in the future. Chasmer and Hopkinson (2016) suggest that the El Niño event in 1998 may have accelerated permafrost thaw. While it has been suggested that rates of permafrost thaw will increase in the future (Jorgenson *et al*., 2011), few studies (*i.e.* work by Baltzer *et al*., 2014 on plateau fragmentation) have identified mechanisms would allow thaw to increase at a rate disproportionate to climate warming. As discussed in chapter four, the presence of shallow taliks substantially increases the rate of permafrost thaw. This research shows that over a five year period (2011 to 2016), the spatial coverage of taliks on peat plateaus increased from 20% to 48%. This rapid increase is attributed to an anomalously warm summer in 2012, suggesting that permafrost thaw can exhibit a non-linear response to climate warming.

In order to successful predict the future availability of freshwater resources in permafrost terrains, it is necessary to correctly parameterize both permafrost thaw rates and the associated changes to basin hydrology. Broadly speaking, this research has provided evidence of: i) increasing permafrost thaw rates through the development of taliks; and ii) increases in hydrological connectivity and stream flows resulting from permafrost thaw.

5.2 FUTURE RESEARCH:

1) This dissertation studies the hydrological processes of peat plateau – bog complexes in discontinuous permafrost terrains. As permafrost continues to thaw, it is predicted that these complexes will continue to disintegrate. Eventually, plateaus will become isolated features in a mosaic of wetlands. In this case, primary runoff pathways will play a much larger role. In this landscape, where permafrost is patchy, further thaw will have a less dramatic effect on changing the contributing area of a basin, but will still affect how water is routed. In a typical peat plateau – bog complex, permafrost impounds water and restricts lateral movement. In a wetland terrain with isolated patches of permafrost, the permafrost is likely to serve two functions: i) displace the water higher in the soil profile; and ii) increase the tortuosity and transit time of water to the basin outlet. These processes may be similar to plot scale hillslope runoff processes in continuous permafrost with mineral hummocks (Quinton *et al*., 2000). A water table located higher in the soil profile may allow for access to zones of the soil with higher transmissivities and therefore increase runoff. This may be countered by longer transit times as water is routed around the permafrost bodies. The response of the system would likely be highly dependent on the per cent cover of permafrost in the basin. Therefore, it is highly unlikely that further permafrost thaw will continue to affect the basin in the same manner as outlined in this

thesis. These hypothesized changes should be evaluated to determine at what point other processes will begin to manifest themselves and exert control on the basin hydrology.

2) Connected taliks, as presented in Chapter 4, offer a conduit for the year-round lateral transfer of mass and energy through a permafrost plateau and have not been previously studied. Most studies examine taliks that are relatively thick (> 2 m) and exist under water bodies or wetlands (Kokelj, Sjöberg *et al*., 2016; McLaughlin and Webster, 2014). There is ample evidence of how the advection energy through taliks can amplify permafrost thaw (Sjöberg *et al*., 2016; de Granpre, 2012), but more work is required to evaluate changes to hydrological connectivity and active layer thickness. Permafrost cored peat plateaus are typically thought to be impervious to subsurface flows given the low hydraulic conductivity of the frozen soil. The presence of a talik enables a year-round subsurface hydrologic pathway through peat plateaus. The hydraulic conductivity in taliks at Scotty Creek is likely very low given the well decomposed nature of the peat (Quinton *et al*., 2008), but the ability for the talik to transmit water year-round suggests that this mechanism should be properly evaluated. Preliminary analysis suggests that the hydraulic gradient between the water levels on either side of a plateau can be quite high for this terrain (*i.e.* 0.05 to 0.1), which could produce non-negligible flow between wetlands and alter contributing areas. Thermal transport associated with water movement through taliks should also be evaluated. The installation of thermistor strings on either side of a connected talik can provide data required to quantify energy transfer year-round. Temperature and soil moisture data throughout the suprapermafrost layer would provide insights into how taliks limit the development of the active layer. Hydrologic and thermal

routing through connected talik networks should be quantified and parameterized to drive spatially distributed hydrologic models in these regions.

3) This research provides process-based information that may serve to parameterize basin scale hydrological models. Scotty Creek, and other adjacent basins are composed of two main land cover areas: 1) the wetland-dominated headwater regions as documented in this study; and 2) forested upland regions underlain by mineral soil. Long-term monitoring meteorological and hydrological data has been collected in the headwater regions, however data has not been collected to explore hydrological processes in the upland regions. Using classified Landsat and IKONOS imagery, Quinton *et al*., (2003) estimate that approximately two-thirds of the Scotty Creek basin is composed of woodlands (combination of peat plateaus and wooded uplands). Unfortunately peat plateaus and uplands were not assigned their own class, but visual observations support the fact that wooded uplands cover a substantial portion of the basin. Geophysical surveys would aid in determining the extent of permafrost in these regions. Traditional methods (*i.e.* frost probing) are difficult as the unconsolidated soil contains stones and rocks, which hinders permafrost detection. Knowledge of the hydrological behaviour of the wooded uplands and the ground ice content is necessary to parameterize basin-scale models.

5.3 POST-PUBLICATION NOTES

This concluding section aims to address any discrepancies between chapters, as each chapter was written separately and may not incorporate insights that were gained subsequently. For example, it was assumed that subsurface interaction between the bogsheds delineated in Chapter 3 (Fig. 3-3) was limited solely to drainage channels. Chapter 4 and Appendix A-1 describe a mechanism whereby subsurface flows may be exchanged through permafrost cored peat plateaus through a talik network. This mechanism was not considered in Chapter 2 and may change the respective contributing areas of the bogsheds. This section aims to address discrepancies between chapters and quantify these errors.

5.3.1 Basin Runoff After 2012

Chapter 2 analysed basin runoff at four Water Survey of Canada gauged basins until 2012. When that chapter was written this was the most recent dataset. At the time of publication of this thesis, runoff data is available until 2015. The two subsequent years after 2012 both experienced very dry conditions, driven primarily by low precipitation. Unadjusted precipitation at Fort Simpson was 329 mm (2013), 245 mm (2014), and 432 mm (2015). Both 2013 and 2014 were below the 30 year average of 388 mm. Most notably, 2014 received very little snowfall (95.6 mm), which was not able to re-wet the soils sufficiently for a below-average rainfall season to produce high amounts of runoff.

Runoff at Scotty Creek after 2012 was 60.4 mm (2013), 22.7 mm (2014), and 115.7 mm (2015). 2013 and 2014 both had very low runoff and do not appear to follow the trend of increasing stream flow as documented in Chapter 2. It is hypothesized that increases to hydrologic connectivity become much more apparent under scenarios of high moisture content. As documented in Chapter 3, when moisture supplies are low, secondary flowpaths are not active. In addition, when the water table is not close to the surface, the saturated zone occupies an area of low transmissivity in the soil, thereby further restricting the subsurface flux. Although flowpaths are more interconnected, this connectivity is not realized when moisture inputs are inadequate to produce runoff at the hillslope or subcatchment scale. When secondary systems are active, a large proportion of the landscape becomes available to produce runoff. Chapter 3 shows that these conditions were rarely met in 2013 and 2014. Under very wet conditions it is expected that secondary runoff may be continuous throughout the entire summer.

5.3.2 Differences between Adjusted and Non-Adjusted Precipitation

There are differences in the precipitation amounts presented in Chapter 2 and those presented in Chapters 3 and 4. Both datasets were from Fort Simpson, NT. The precipitation amounts in Chapter 2 are from the Adjusted and Homogenized Precipitation dataset from Environment Canada. Precipitation values presented in Chapters 3 and 4 are unadjusted values from Environment Canada.

After initially using the adjusted dataset for Chapter 2, we reverted to the unadjusted dataset for the two subsequent papers. The adjusted dataset showed a significant trend of increasing precipitation, mainly in the form of snowfall, over the past 30 years. In the last 15 years of available data (1998-2012), average annual adjusted snowfall was 262 mm, with three years exceeding 300 mm. To date, the largest end of season snowpack that has been measured at the SCRB by means of snow surveys was 172 mm, with an average of 117 mm over 12 years. Although it is expected that there will be some loss of snow between cumulative daily snowfall and end of season snowpack, the adjusted values were more than double what was measured on the ground. The average unadjusted SWE at Fort Simpson over the same time period was 153 mm with a maximum of 191 mm. These values are much closer to what was observed on the ground in the SCRB. Interestingly, the unadjusted data does not display a significant positive trend increase over the same 30 year period. GlobSnow, a remote sensing product used to measure SWE, also does not report an increasing trend in snowfall in the Scotty Creek region over the last 30 years. This indicates that there may be an artificial inflation in SWE in the adjusted dataset. For this reason, we avoided using the adjusted dataset in subsequent chapters.

5.3.3 Changes in Subsurface Connectivity Through Talik Networks

A key assumption in Chapter 3 was that the permafrost cores of peat plateaus restrict subsurface transmission of water due to the very low hydraulic conductivity of the frozen substrate. This chapter failed to incorporate new insights that were gained while conducting research for Chapter 4. A residually thawed layer (talik) between the active layer above and the permafrost below can allow for year-round transmission of water, thereby fundamentally changing the contributing areas for individual bogsheds. Field work has shown that connected taliks typically only exist when the plateau width between two wetlands is < 30m. In the bog cascades studied in Chapter 3, there was only one instance where a connected talik was found between wetlands. This talik connected the most upstream bog in the west cascade (W-1) with a bog to the south (in a different, unstudied cascade). Due to the relatively small cross sectional area of the connected talik, relative to the size of the catchment (west cascade), it is hypothesized that the change in contributing area is minimal and will not significantly affect the runoff or runoff ratio measurements made in Chapter 3. Furthermore, Chapter 3 focussed primarily on surface runoff, as subsurface transmission in the talik zone of the drainage channel would be orders of magnitude smaller. In light of these new flowpaths, ongoing research is being conducted to quantify how taliks transmit water and energy and how contributing areas may expand/contract in response to these flowpaths.

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APPENDIX A-1:
Active layer and talik dynamics of a permafrost cored peat plateau

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ABSTRACT:

The presence of a talik (perennially thawed feature in a permafrost environment) below the active layer (ground that freezes and thaws annually) has important hydrologic and thermal implications as taliks can provide an active flowpath throughout the year. As taliks do not freeze over winter, they should be distinguished separately from the active layer. In areas of discontinuous permafrost where taliks are prevalent, measuring end of season thaw depth should not be considered a proxy for active layer thickness. At a site in the southern Northwest Territories, Canada, we measured active layer thickness and talik thickness across an ice-cored peat plateau in 2015 and 2016. In both years the average thickness of the active layer was 35 cm and 41 cm respectively, whereas talik thickness was 80 cm and 99 cm. The talik extended the entire width of the plateau and provides a year-round hydrological connection between a flat bog and a channel fen. The presence of a talik will increase the rate of permafrost thaw as it allows for greater advection of energy through this feature. As the active layer thaws in the spring, a two-layered flow system develops as snowmelt water flows over the frozen portion of the active layer of plateaus into adjacent wetlands, and stored water from the plateau is also routed through the unfrozen, saturated talik. This two-layered system continues until the entire active layer thaws. Flow through the talik is limited by the low hydraulic conductivity of deep peat (~ 3.5 m day^{-1}), but as it persists throughout the year it amounts to ~10% of total runoff from the plateau.

INTRODUCTION:

The active layer, defined as the 'top layer of ground subject to annual thawing and freezing in areas underlain by permafrost' (ACGR, 1988), is conceptualized as the layer of soil where most hydrological processes occur. Runoff typically occurs in the thawed portion of the active layer, above the underlying frozen ground (Carey and Woo, 1999; Wright *et al*., 2008) as the saturated hydraulic conductivity of unfrozen soil is orders of magnitude higher than frozen soil. As the majority of hydrological processes occur within the active layer, changes in active layer thickness will have important hydrological consequences, and thus should be monitored (see Brown *et al*., 2008).

In areas of permafrost with high ice content, a large amount of energy is required to satisfy latent heat requirements to thaw the frozen active layer. Conversely, the ground must also lose an equivalent amount of energy to re-freeze the active layer over winter. The amount of energy the ground loses over-winter is controlled primarily by the timing and magnitude of snowfall, as snow is a very effective thermal insulator (Williams and Smith, 1989). The thickness of the active layer is thus governed by whichever process (thawing or freezing) is least able to penetrate the ground. In areas of continuous permafrost, seasonally thawed ground typically refreezes entirely, and consequently, the maximum thaw depth defines the active layer thickness. However, in areas where the mean annual temperature approaches 0°C, it is necessary to measure both freezing and thawing depths when determining active layer thickness. If insufficient energy is lost over winter such that a complete refreezing of the active layer does not occur, a talik forms between the active layer and the underlying permafrost. Depending on soil moisture conditions and water table location, this layer may or may not be saturated. A series of consecutive warm summers and/or winters may allow the talik to grow to a thickness such that complete refreezing of the soil above permafrost may not be possible over one winter.

Accordingly, we caution that measurements of maximum (*i.e.* end of summer) thaw depth should not be taken as the active layer thickness. The Circumpolar Active Layer Monitoring (CALM) program protocol states that active layer thickness can be measured by 'late-season mechanical probing' of the ground (Brown *et al*., 2008, p. 169). This incorrectly assumes that the entire depth of ground above permafrost completely refreezes over winter, an assumption that yields errors in energy balance calculations, particularly concerning phase changes. This assumption also incorrectly implies that hydrological processes are relatively dormant for a period of time when the ground is assumed to be entirely frozen and does not account for the possibility of a talik. Unfortunately, the freezing depth is much more difficult to measure than the thaw depth, and as such it is rarely reported.

When S.W. Muller (1947) first presented the term 'active layer', he also proposed that the term '*suprapermafrost layer*' be used to describe the 'combined thickness of ground above the permafrost consisting of the active layer and talik' (Muller, 1947 p. 11). We recommend that the term 'suprapermafrost layer thickness' be used in place of 'active layer thickness' when depth to permafrost table measurements are taken at the end of the thaw season, but not corroborated with measurements of the maximum penetration of the freezing front.

Permafrost cored features such as peat plateaus are typically thought to inhibit the transmission of subsurface water beneath the thawed portion of the active layer due to the very low hydraulic conductivity of frozen, saturated peat. The current conceptual understanding of hillslope runoff from these plateaus is that runoff only occurs in the thawed portion of the active layer (Wright *et al*., 2008). The presence of a talik may provide an additional runoff pathway on hillslopes and allow for the transmission of water between wetlands throughout the year. The objectives of this study are two-fold: 1) to document the hydrological function of a perennially-thawed talik; and 2) to show that the thickness of the active layer at our study site is governed by the depth of re-freeze over winter and not by the maximum summer thaw depth.

STUDY SITE:

The study was conducted at the Scotty Creek Research Basin (SCRB), located about 50 km south of Fort Simpson, NT (Figure 1a). The study site is dominated by thick peat deposits (>2 m) overlying a clay rich glacial till of low hydraulic conductivity. Permafrost occupies ~40% of the basin and exclusively takes the form of treed peat plateaus that rise about 1m above surrounding wetlands (channel fens and flat bogs). The associated hydraulic gradient directs runoff from the plateaus and into the adjacent wetlands (Wright *et al*., 2008). Channel fens transmit water to the basin outlet (Hayashi *et al*., 2004), while flat bogs can either act as storage features or route water to the channel fens through a fill-and-spill process dependent on antecedent moisture conditions (Connon *et al*., 2015). Ongoing data collection has been occurring since 1999 at a study plateau. The plateau is flanked by a channel fen on one side and a flat bog on the other (Figure 1b). Substantial lateral and vertical thawing of permafrost at the study plateau has been observed since monitoring began (Figure 2).

METHODS:

End of summer (late August) measurements of permafrost table depth have been taken at the study plateau since 1999. These measurements have been taken annually along a transect at 1 m intervals, permitting 19 measurement points in 2014 and 2015. This allows for measurement of both lateral and vertical permafrost thaw. In 2015, weekly measurements of the depth to the frost table (top boundary of the frozen and saturated soil) were taken to quantify the progression of thaw. Typical downward progression of the frost table averages about 0.5-1 cm day^{-1}. When a sudden increase (*i.e.* more than 30 cm) in depth to frost table was observed it was assumed that this was the boundary of seasonal re-freeze and that the measured frost table was now below a talik. In 2016, end of winter (early April) measurements of maximum re-freeze depth were taken by using a hand ice auger to drill through the frozen soil until the unfrozen talik was reached. The boundary between frozen and unfrozen soil was clear and this depth was measured and recorded as active layer thickness.

Thermistors and water content meters are installed at 10 cm increments in the soil to quantify temperature and liquid water content at different depths. Additionally, in 2016, the temperature of the talik was also measured using a handheld digital thermometer. Total pressure transducers were installed in the bog and fen adjacent to the plateau to calculate a hydraulic gradient between the two features. The hydraulic conductivity of the peat in the talik is assumed to be 3.5 m day^{-1} (Quinton *et al*., 2008, M. Braverman unpublished data). Although there was only one transect, it is assumed that the talik thickness calculated on the transect is representative of the entire 75 m length of the plateau. Total flux through the plateau was then calculated using Darcy's law.

RESULTS:

The average thaw depth measured at the end of summer was 115 cm in 2014 and 135 cm in 2015, while for the same points, the average refreeze depth was 35 cm in 2015 and 41 cm in 2016. In both years, the thaw depth was ~3 times greater than the depth of re-freeze, indicating a talik with a thickness of ~1 m (Fig. 2). The thickness of the talik and has increased each year since measurements began, while the width of the plateau has decreased as the permafrost core thaws laterally (Fig. 3). Ground thaw, driven by the vertical heat flux from the ground surface, is augmented by advection of energy via water moving from the bog to the fen. Both continuous and discrete temperature measurements indicate that the talik is isothermal at -0.2°C, the freezing point depression measured at this site (Quinton and Baltzer, 2013). Liquid volumetric soil moisture at 50 cm depth (deepest soil moisture sensor) was 0.8, indicating that the soil was fully saturated with liquid water throughout winter. The total flux of water draining through the talik and into the fen is 47 mm yr^{-1}, accounting for about 10% of total average runoff (520 mm yr^{-1}) from the plateau (data from Quinton and Baltzer, 2013).

Given the thickness of the talik, it is highly unlikely that enough energy could be removed from the suprapermafrost layer to freeze it entirely. Given current climate conditions in the study region, once a talik expands vertically to the point where complete refreezing in winter is no longer possible, thaw of the underlying permafrost is inevitable, owing to the presence of liquid water on the permafrost table throughout the year.

DISCUSSION:

Hillslope runoff from a permafrost cored peat plateau as first described by Wright *et al*. (2008) indicates that runoff is restricted to the thawed portion of the active layer. Combining the results of the current study with that of Wright *et al*. (2008) suggests a two-layered runoff system where both the talik and thawed portion of the active layer convey subsurface runoff. Although not the primary runoff mechanism, flow through the talik should not be excluded from runoff measurements. Using isotope tracers, Hayashi *et al*. (2008) found that less than half the runoff from the SCRB was derived from snowmelt (event) water. Other studies (*i.e.* Gibson *et al*., 1993; Carey *et al*., 2012) using isotope tracers have found similar results in discontinuous permafrost terrains. These studies found that 'old' water dominated the hydrographs; the presence of taliks would provide a flowpath allowing for this old water to reach the drainage network.

Many studies (*i.e.* Akerman and Johansson, 2008; Xue *et al*., 2008) have reported a trend of a thickening active layer in climates where the mean annual temperature is close to 0°C but do not report the depth of refreeze. It is important that researchers do not incorrectly assume that the thaw depth measured at the end of the thaw period is a measure of the thickness of the active layer. We show this assumption to be erroneous, and it is important to distinguish the thickness of the suprapermafrost layer and the thickness of the active layer if the suprapermafrost layer does not entirely re-freeze over winter. Therefore, researchers should be cognoscente of the fact that end of season thaw depths may not be indicative of active layer thickness (Muller, 1947; ACGR, 1988), especially in areas of discontinuous and sporadic permafrost.

We suggest that climate warming leads not to active layer thickening as commonly discussed in the literature, but to active layer thinning, considering that the largest temperature increases in Canada's North have occurred over winter (Vincent *et al*., 2015). A thinner active layer would allow for more rapid thawing of the underlying permafrost, as thawing of the active layer would be completed earlier in the season, permitting more energy to penetrate through to the permafrost. These processes must be properly conceptualized and parameterized in order to better understand and predict the response of such systems to climate warming.

Permafrost thaw is very rapid in the zone of discontinuous permafrost (Kwong and Gan, 1994; Quinton *et al*., 2011) where permafrost is thin (<10m), relatively warm (>-1.5°C), and subject to both lateral and vertical thaw. Permafrost thaw changes the routing and storage of water within a basin (St. Jacques and Sauchyn, 2009; Connon *et al*., 2014). If a thinner active layer facilitates more rapid permafrost thaw, the coupled hydrological changes will also occur more rapidly than expected. It is important to identify the point at which thinning of the active layer may be expected, as this would indicate a threshold at which permafrost thaw will become more rapid.

CONCLUSIONS:

In permafrost regions where taliks have developed, active layer thickness should be measured by the maximum extent of the freezing front, not by the thaw depth at the end of summer. In these areas, a thinning of the active layer is predicted in response to a warming climate. We propose that the term '*suprapermafrost layer*' originally suggested by Muller (1947) be reincorporated into the current nomenclature to refer to the end of summer thaw depths when re-freeze depths are not known. The presence of a talik provides an additional flowpath that can transport water to the drainage network year-round. Taliks also allow for lateral advection of energy, providing an additional energy source that may thaw underlying permafrost, and as such should be documented and included in runoff measurements and models to accurately represent the system.

FIGURES:

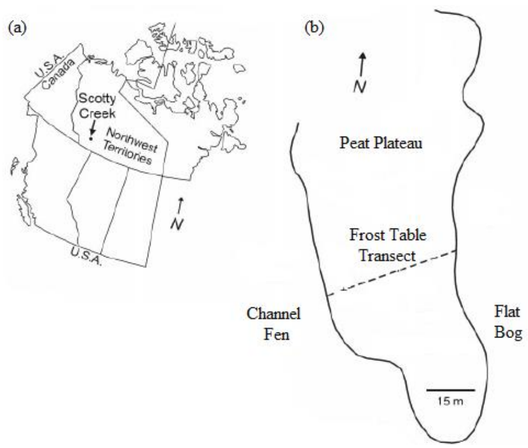

Figure 1: (a) Location of SCRB; (b) planar view of study plateau. Modified from Wright *et al.* (2008)

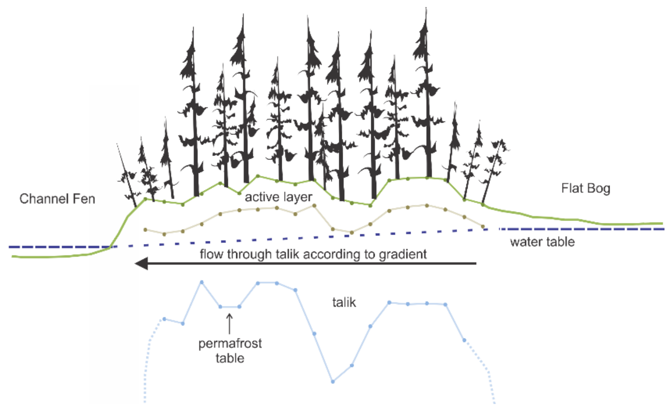

Figure 2: Cross section of study plateau indicating thickness of the active layer in and underlying talik. Each dot indicates measurement location. Data shown is from 2014/2015.

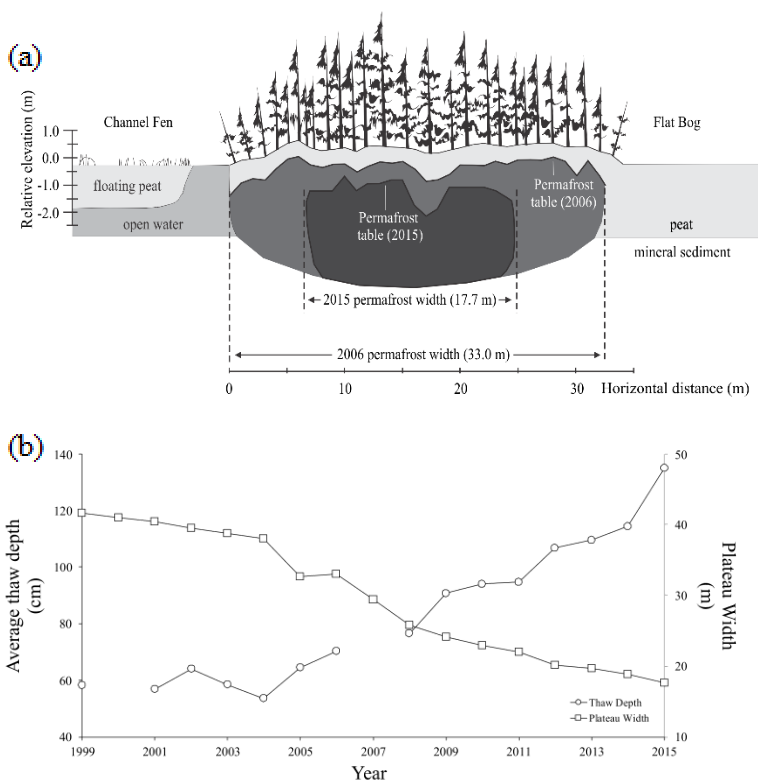

Figure 3: (a) Cross section of study plateau showing lateral and vertical thaw since 2006;
(b) Changes in plateau width and thaw depth since 1999

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