WILDFIRE IMPACTS ON PEATLAND ECOHYDROLOGY

WILDFIRE IMPACTS ON PEATLAND ECOHYDROLOGY

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ABSTRACT

Forested peatlands of the Canadian boreal forest represent a globally significant carbon pool subject to periodic disturbance from wildfire. The current ecohydrological conceptualization of peatland recovery from (and resilience to) disturbance is largely based around work on cutover peatlands. Wildfire disturbance is simultaneously more spatially extensive and more variable in the severity of disturbance compared to cutover peatlands. The objective of this thesis is to examine the changes to peatland ecohydrological processes as a result of wildfire disturbance in forested ombrotrophic peatlands of the Boreal Plains. The hydrology and atmospheric exchanges of energy and water were examined at two peatlands in northern Alberta: one recently burned and the other approximately 75 years since fire.

Wildfire resulted in little change in net radiation flux to the peatland during the snow-free period. A decrease in the net radiation flux during the late winter was caused by the loss of the tree canopy and the increase in albedo during winter. While summer albedo largely returned to pre-fire values within two years after fire, the amount of solar radiation reaching the burned peat surface increased by nearly 50%. As a result, surface evaporation increased by an amount only marginally greater than the loss of transpiration. The net result on the water balance was a modest increase in water losses during the course of the summer, resulting in a lower water table. Water table decline per unit of evaporation was higher due to a decrease in specific yield, likely from a combination of post-fire peat compression and the combustion of high specific yield surface peat during wildfire. The combination of lower water table and enhanced evaporation cause greater pore-water pressures after fire, particularly in hummocks. The hydrological regime of hollows was not significantly altered by wildfire, despite the larger depth of burn in the hollows.

ACKNOWLEDGEMENTS

I would like to thank my supervisor Dr. Mike Waddington for his support and guidance through the years on this project. Without his careful persuasion during a conversation at the Bois-des-Bel peatland, Quebec, in June 2007, none of this work would have ever come to fruition. His patience in dealing with my ever changing interests is not to be underestimated. Much of this work is based on the ecological studies of Drs. Merritt Turetsky and Brian Benscoter. Without their guidance, in particular for introducing me to the peatlands of Alberta, this work would not qualify under the banner of ecohydrological studies. I am in debt to the entire PEATFIRE research group, including those mentioned above, for fostering such a healthy and interdisciplinary research environment.

My very entry into graduate studies would not have occurred without the support of Drs. Kathy Young and Ming-ko Woo, who first hired me as a wet behind the ears research assistant to traverse the wetlands of the High Arctic in 2004 and 2005.

My family, friends, and loved ones were instrumental in their support and patience during my absences of up to six months of the year. I was always grateful to have such people to look forward to seeing again after my long absences in the field.

Over twelve months of total fieldwork were involved this work, the majority of which was done with the field assistance of Steve Baisley and James Sherwood. I am glad to have them as friends and coworkers; I am happy to see their own research building on the work shown here. The other PEATFIRE students and field assistants are too numerous to mention, but I am grateful to them all. In Alberta, the logistical and leisure support of the Meanook Biological Station was instrumental in making each field season a memorable one.

This work was supported via a NSERC Alexander Graham Bell CGS-D scholarship, and a NSERC strategic grant to the PEATFIRE project. I am grateful to the Canadian Forest Service, Northern Forestry Centre, for hiring me as a research scientist during the final months of the writing process.

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DECLARATION OF ACADEMIC ACHIEVEMENT

This dissertation takes the form of a series of manuscripts to be submitted to various academic journals. There is some degree of repetition in the site description and methods. Latter chapters do in some degree build on the previous ones, but only by utilizing some minor data. Each of the chapters can be read as a stand-alone document. The reference style in the text has been standardized for the purposes of this dissertation, but the citations at the end of each chapter remain in the formatting of the journal of submission. Otherwise the text and figures appear in the form that will be submitted for publication in journals.

In addition to the work presented here, I authored or contributed to four other peer-reviewed publications relevant to this thesis in the course of my graduate studies.

A review of *Sphagnum* and the vadose zone processes controlling its carbon exchange and photosynthesis came about from an assignment in a graduate course:

Thompson, D.K., and Waddington, J.M. (2008), Sphagnum under pressure: towards an ecohydrological approach to examining *Sphagnum* productivity. *Ecohydrology, 1,* 299-308.

A workshop on peatland geophysics hosted by the University of Maine at Orno resulted in a paper on the utility of geophysical methods in detecting subsurface structures in peatlands. I played a minor role in the fieldwork and analysis of the resistivity profiles:

Kettridge N., Comas, X., Baird, A., Slater, L., Strack, M., Thompson, D.K., Jol, H., and Binley, A. (2008), Ecohydrologically important subsurface structures in peatlands revealed by ground-penetrating radar and complex conductivity surveys. *Journal of Geophysical Research, 113*, G04030, doi:10.1029/2008JG000787.

After installing instrumentation at the study sites in the fall of 2007, I participated in a series of laboratory experiments in the dynamics of peat combustion with Brian Benscoter. I contributed significantly to the eventual publication of the results by creating a model of peat combustion that was successfully implement using data from the lab experiments:

Benscoter B.W., Thompson, D.K., Waddington, J.M., Flannigan, M.D., Wotton, B.M. DeGroot, W.J., and Turetsky, M.R. (2011), Interactive effects of vegetation, soil

moisture, and bulk density on the depth of burning of thick organic soils. *International Journal of Wildland Fire, 20,* 418-429.

During the course of my research assistantship, I contributed to a study examining the relationship of peatland water tables to the drought code of the Canadian Forest Fire Danger Rating System. I contributed the analysis for the paper, alongside two years of water table data:

Waddington, J.M., Thompson, D.K., Wotton, M., Quinton, W.L., Flannigan, M.D., Benscoter, B.W., Baisley, S.A., and Turetsky, M.R. (2012), Examining the utility of the Canadian Forest Fire Weather Index System in boreal peatlands. *Canadian Journal of Forest Research, 42*, 47-58. doi: 10.1139/x11-162.

The work presented in this thesis is the result of collaborative research, and the specific contributions of the candidate are listed below.

Chapter 2

Title: Effect of wildfire on forested peatland albedo and radiation exchange

Authorship: D.K. Thompson A.S. Baisley, M.R. Turetsky, and J.M. Waddington

Status: To be submitted to *Journal of Geophysical Research*

Candidate's contribution: All data except shrub LAI and below-canopy albedo measurements were collected and analyzed by the candidate. Manuscript preparation was completed by the candidate; J.M. Waddington provided feedback to the manuscript. M.R. Turetsky and J.M. Waddington secured funding for the project.

Chapter 3

Title: Contrasting the water balance of a burned and unburned forested peatland

Authorship: D.K. Thompson, J.M. Waddington, B.W. Benscoter, and M.R. Turetsky

Status: To be submitted to *Hydrological Processes*

Candidate's contribution: B.W. Benscoter assisted in the collection of the evaporation measurements. The candidate collected the rest of the data and conducted the analysis and manuscript preparation. J.M. Waddington provided feedback to the manuscript. M.R. Turetsky and J.M. Waddington secured funding for the project.

Chapter 4

Title: Hydrophysical properties of peat in forested peatlands subject to wildfire disturbance

Authorship: D.K. Thompson and J.M. Waddington

Status: To be submitted to *Water Resources Research*

Candidate's contribution: The candidate conducted the lab analyses and manuscript preparation. J.M. Waddington provided comments on the manuscript and funding.

Chapter 5

Title: Impact of wildfire on the vadose zone hydrology of a forested boreal peatland

Authorship: D.K. Thompson and J.M. Waddington

Status: To be submitted to *Journal of Hydrology*

Candidate's contribution: The candidate conducted the fieldwork and analyses, as well as the manuscript preparation. J.M. Waddington provided comments on the manuscript and funding.

CHAPTER 1: INTRODUCTION

1.1. Disturbances in Northern Peatlands

Peatlands store approximately 455 Pg of carbon globally (Gorham, 1991). In Canada, peatlands cover 1.1 x 10^4 km^2 (Tarnocai *et al.*, 2006). The largest disturbance on peatlands is wildfires, releasing 4700 Gg C a^{-1} over ~1500 km^2 annually in western Canada alone (Turetsky *et al.*, 2002). This annual area of peatlands disturbed by wildfire far exceeds industrial peatland disturbances such as horticultural extraction, hydroelectric development, or hydrocarbon extraction (Turetsky *et al*., 2002), but to date has received relatively little attention from a process ecohydrology perspective.

In the western boreal forest of Canada, the annual area burned of peatlands is a roughly fixed proportion of the total landscape, reflecting how peatlands contribute to the mosaic of burned area equally alongside uplands (Turetsky *et al.,* 2002). In Alaskan black spruce lowlands, Shetler *et al.* (2008) demonstrated how thick hummock microfroms of *Sphagnum fuscum* ("*Sphagnum* sheep" *sensu* D.H. Vitt) serve to retard the combustion of the peat surface compared to other mosses and lichens. In a laboratory experiment, Benscoter *et al.* (2011) showed *S. fuscum* was particularly resistant to combustion and does not contain sufficient bulk density to contribute to combustion and rather serves as an energy sink even at water contents observed during deep water table (*WT*) drawdown. Benscoter and Vitt (2008) showed how this key peat-protecting species reaches a maximum abundance 40 years after burn and decreases subsequently as more combustible feathermosses propagate as the spruce canopy closes. Thus, there is a need to systematically couple the previously detached processes of peat combustion and ecosystem ecology via the common framework of ecohydrology, specifically the growth and water retention of keystone peat-forming species such as *Sphagnum fuscum*.

In boreal upland forest fires, fuel consumption, surface albedo, vegetation succession, and fire return period determine the long-term radiative forcing from a wildfire (Randerson *et al.,* 2006). Hydrology is a major control on vegetation succession in peatlands, where the key peat-forming species, which lack vascular systems, rely entirely on a favourable near-surface hydrological regime. Given that vegetation succession ecology is a function of fire severity and the ability of vegetation to recolonize the burned substrate (Benscoter, 2006), it is likely that hydrological studies will provide a vital knowledge on the recovery of peatland ecosystems from fire disturbance. However, to date no studies of post-fire peatland hydrology have been undertaken to complement existing ecological and carbon biogeochemistry studies.

1.2. Ecohydrology of Forested Peatlands

1.2.1. Surface Energy Balance

Studies of vegetation succession points to a return to a pre-fire cover of *Sphagnum* within 25-40 years after fire (Benscoter and Vitt, 2008), while microclimatological observations along upland boreal forest chronosequences (Amiro *et al*., 2006) show that 50-75 years is required before pre-fire energy balances are achieved. Given the slower growth of black spruce in lowland forests and peatlands (Bond-

Lamberty *et al*., 2004) but sparse tree canopy, it is likely the return to a pre-fire energy balance in peatlands will be delayed relative to uplands.

In unburned conifer stands, mid and late winter albedo is low due to snow sublimation and ablation from the canopy (Arain *et al*., 2003). The loss of tree cover reduces snow sublimation thereby increasing winter-time albedo (Liu *et al.,* 2005; Randerson *et al.,* 2006). Snow-free albedo in peatlands generally is lowest before shrub leaf-out (Lafleur *et al*., 1997) with a maximum in late summer just before senescence. Field observations (Shetler *et al.*, 2008) and lab ignition experiments (Benscoter *et al.*, 2011) show that these *Sphagnum* "sheep" are created after the loss of the capitula and the upper few centimetres of the moss layer. Any remaining charcoal and ash is quickly eroded and transported into the lower-lying hummocks (Turetsky, pers. comm.), leaving a white-coloured microform with a high albedo. This reflective surface in hummocks may reduce cumulative net radiation and ground heat flux. Hummocks may also undergo an additional albedo alteration during the summer where the albedo of *Sphagnum* sheep may increase as volumetric water content (*VWC*) decreases during summer. This pattern of albedo being negatively correlated with *VWC* has been observed in peatlands by McMorrow *et al*. (2003). The opposite albedo regime potentially exists in hollow microforms, where the dark coloured ash and char rich substrate may increase net radiation and ground heat flux.

In an investigation of the summer radiation balance of a burned taiga stand, Rouse (1976) found that despite a 73% decrease in albedo, the dry surface crust of ash (small, mobile elemental carbon particle) and charcoal (incompletely combusted organic carbon) emitted more longwave radiation, causing a small decrease in net radiation at the surface. In summer, albedo is very spatially heterogeneous between hummocks and hollows, but an overall small decrease in albedo is anticipated as a function of time since fire, as seen in forested uplands (Amiro *et al.,* 2006).

Alongside large changes in albedo, removal of the tree canopy has also been found to increase the proportion of net radiation allocated to the ground heat flux. Increases in the ground of heat fluxes in recently harvested aspen stands of 220% (Amiro, 2001) and in lowland spruce stands of 30% (Chambers *et al.*, 2005) are in both cases attributable to an increase in incident radiation reaching the soil surface along with decreased albedo. The increase in ground heat flux makes the soil surface more responsive to net radiation variation through the day, leading to larger diurnal fluctuations in near-surface soil temperature (Amiro, 2001; Chambers *et al*., 2005), though this response may be tempered by a higher *WT* and increased near-surface moisture content.

1.2.2. Water Balance

Numerous field investigations and modelling studies (*e.g.* Amiro, 2001; Bond-Lamerty *et al*., 2009) have pointed to a drastic reduction in overall evapotranspiration immediately after wildfire in upland boreal forest stands. While a similar reduction in transpiration can be hypothesized from the loss of the tree canopy, it is unknown how the uniquely high storage capacity of deep peat deposits reacts to a sudden decrease in transpiration. The decrease in transpiration from wildfire may potentially cause more water to be stored instead of leading to increased runoff, as is the case in forested hill slopes affected by fire (Inbar *et al*., 1998).

In unburned peatlands, previous examinations have shown trees and shrubs play a major role in peatland hydrology by simultaneously lowering surface moisture by transpiration and interception (van Seters and Price, 2002) and providing shade that reduces surface evaporative demand on the moss surface. In the absence of fire, the tree and shrub layers are able to access moisture closer to the *WT* at rates in excess of capillarity, maintaining evaporation even at low *WT* levels (Admiral and Lafleur, 2007). Moreover, the abrupt removal of trees from forested peatlands during logging raises the water table (Simard *et al.*, 1997), but the water table response may also be influenced by changes in surface evaporation and runoff (Verry, 1997). In upland conifer stands, Amiro (2001) found a 50-75% decrease in the latent heat flux one year after wildfire in a jack pine stand in Northwest Territories without any significant change in net radiation. While a mature jack pine stand has a basal area and leaf area index (*LAI*) far greater than a forested black spruce bog, a similar response from loss of transpiration is anticipated, though the relative contribution of transpiration in the water balance may differ.

In the few studies of surface energy and water exchange at the microform scale, Admiral and Lafleur (2007) found that evaporation in hummock microforms from a shrubby but treeless bog in eastern Ontario is often moisture-limited during dry periods, leading to an increase in the sensible heat flux. This increased latent heat flux then causes microadvective increases in evaporation at adjacent hollows as warm and dry air passes over them. Similarly, the combustion of the surface capitula layer and loss of shrub canopy potentially allows for a higher maximum vapour conduction from the unsaturated zone and increases the maximum evaporation rate.

Process hydrological and climatic examinations of microtopography in peatlands are limited. Both Kellner (2001) and Admiral and Lafleur (2007) observed enhanced evaporation from hollows as a result of microadvection from drier hummocks to more wet hollows at the Mer Bleue bog. In observation of chamber evaporation in lowland black spruce forests forests, Heijmans *et al.* (2004) found that surface evaporation rates were similar across *Sphagnum*, true mosses, and shrubs, but lichens showed a significantly lower evaporation rate. Recently, Bond-Lamberty *et al.* (2011) showed that a diverse range of feathermosses and *Sphagnum* mosses show a similar increase in surface resistance to evaporation with drying (as measured by gravimetric water content). However, feathermosses showed significantly lower evaporation rates in the field. This contradiction suggests the importance of the surface energy balance in determining evaporation rates (and controls on moss moisture itself) rather than on a species-controlled evaporation rate dictated by surface resistance values.

1.2.3. Hydrophysical Peat Properties

Alterations in hydrophysical properties in cutover peatlands have been found to be major drivers of changes in hydrological cycling and energy balance (Price *et al.,* 1998). Specifically, the properties of peat in the unsaturated zone alter the potential storage and transmission of water from the *WT* to the surface to satisfy evaporative demand. In circumstances where the upward movement of water from the saturated zone to the atmosphere is insufficient compared to evaporative demand, preferential drying at the surface may result, leading to low *VWC* and greater pore-water pressures near the surface (Price, 1998). While such equilibrium conditions are commonplace in cutover peatlands (Lindholm and Marrkula, 1984), their occurrence in peatlands subject to wildfire is unknown.

Knowledge of the pore-size distribution along a vertical peat profile provides insight into the equilibrium surface *VWC* as a function of *WT*, and therefore horizontal and vertical distribution of vegetation communities. Using a 1-D implementation of the Richards equation, the Hydrus model of Simunek *et al.* (2008) can resolve an equilibrium *VWC* profile given a constant atmospheric evaporative demand and *WT*. This equilibrium *VWC* profile is a useful tool for comparing profile hydrophysical characteristics across sites where other factors such as climate and individual site characteristics do not provide an independent comparison of surface *VWC* and pore-water pressure. While some empirical relationships between peat properties and the Van Genuchten (1984) model used to drive Hydrus are known, the observations are limited to deeper horizons in drained peatlands. The behaviour of shallow, low density peat as well as the role of coarse roots and woody content in altering peat water retention properties is currently unknown. While peat properties under drainage disturbance are well-understood (Schlotzhaur and Price, 1999), the impact of wildfire on peat physical characteristics is less constrained. Mallik *et al.* (1984) noted that pore size distributions shifted towards a smaller mean pore size after fire in British heather-dominated peatlands, even in horizons not directed exposed to flaming or smouldering. It was hypothesized that ash (small particles of elemental carbon) residue washes into lower soil horizons, clogging large pores. A reduction in mean pore size will likely increase water retention, leading to higher soil moisture than would be found in the absence of clogged pores. Whether such alterations also take place in *Sphagnum*-dominated systems is unknown.

1.2.4. Unsaturated Zone

In maritime peatland systems the differentiation between microtopography occurs at the hummock-pool level, where *Sphagnum* is differentiated from open-water pools (Rietkerk *et al*., 2006), commonly referred to “string and flark” topography (Rydin and Jeglum, 2006). In some cases, low-lying topography is split into vegetated flat “lawns” (*c.f.* Strack *et al.*, 2006) and open-water pools. In the western boreal forest of Canada, open-water pools are not generally present in bogs, and “hollows” are instead populated primarily by a mixture of the hollow by *Sphagnum angustifolium*, lichens (e.g. *Cladina* and *Cladonia spp*.), and feathermosses (e.g. *Pleurozium spp*.). While lichens are often associated with dry habitats, these low-lying hollows often experience extremely dry soil moisture conditions during deep water-table depths in summer (Rydin, 1985). In contrast to the varying soil moisture in hollow microforms, hummock microforms dominated by *Sphagnum* species similar to *S. fuscum* have been observed to remain relatively more moist even a *WT* greater than 50 cm below the surface (Rydin, 1985).

In undisturbed peatlands, *WT* position is commonly used as a broad predictor of moisture availability at the surface in models and field studies (*e.g.* Hayward and Clymo, 1983; Rydin, 1985; Li *et al.,* 1992; Silvola, 1996; Bubier *et al.,* 2003). In peatlands affected by fire, the *WT* dynamics may become detached from the surface soil moisture, a behaviour observed in cutover peatlands by Price (1997). As such, the direct measurement of soil moisture in the form of *VWC* or pore-water pressure is required to overcome uncertainties of the *WT*-surface moisture relationship in disturbed peatlands. While *VWC* is the simplest and most common direct measure of surface moisture, it suffers from variable and non-stationary relationships to physiological thresholds for *Sphagnum* growth and peat formation. In contrast to *VWC*, measurement of pore-water pressure, is a more intensive measure that corresponds to the capillary capacity of living mosses (Hayward and Clymo, 1982), and hence provides a first-order indicator of surface moisture as it relates to moss growth (Thompson and Waddington, 2008).

The structure of peat plays an important role not only in the largely lateral flow in the saturated zone, but also in water retention and capillarity in the unsaturated zone. Poikilohydric plants such as *Sphagnum* mosses rely of capillary forces to transport water from the saturated zone to the surface (Proctor, 2000). Surface pore-water pressure is a more accurate, though rarely measured, hydrological state that is sensitive to capillarity and the water retention abilities of mosses, as it can be related to the same fundamental structure of water allocation across a pore-size distribution (Hayward and Clymo, 1982; *c.f.* Thompson and Waddington, 2008). Hayward and Clymo (1982) calculated the drainage of hyaline cells – a critical water retention structure in *Sphagnum* – to occur at a range of pore-water pressures from -100 to -300 mb. In field investigations, Price and Whitehead (2001) observed that *Sphagnum* will not establish in environments where the minimum pore-water pressure exceeds -100 mb at any point in the summer.

Although no data exists for peatlands affected by wildfire, studies from harvested peatlands show that *Sphagnum* colonies are more sensitive to fluctuations in *VWC* and *WT* compared to undisturbed peatlands (Tuittila *et al.,* 2004). In studies at the Cacouna bog in Quebec, regenerated *Sphagnum* hummocks 30 years since harvesting showed a similar robustness to hydrological stress compared to natural peatlands (Thompson and Waddington, unpublished data). While the mechanism of disturbance in drained and cutover peatlands is fundamentally different, the physiological constraints on *Sphagnum* colonization, namely moisture availability, may be similar. It is hypothesized that thresholds of pore-water pressure circa -100 mb observed in field studies (Price and Whitehead, 2001) and laboratory studies (Hayward and Clymo, 1982) will be comparable to hydrological thresholds after wildfire because of the fundamentally similar restriction of sufficient capillary movement of water within the uppermost living *Sphagnum* layers.

Hummock species, such as *Sphagnum fuscum*, have been shown to maintain higher water contents as compared to low-hummock or hollow species at equally deep water tables (~ 20 cm) (Rydin, 1985). Titus *et al.* (1983) hypothesized the higher water retention capacity of hummock microform mosses is related to dessication avoidance, while hollow microform species such as *Sphagnum angustifolium* are more tolerant of dessication. Shipperges and Rydin (1998) emphasized the importance of the capillarity network and dense surface capitula of *Sphagnum* in maintaining high water contents and did not otherwise find any species specific difference in dessication resilience. In any case, it appears the physiology of hummock species such as *Sphagnum fuscum* is especially able to maintain sufficient moisture for growth even during periods of typically deep water tables during summer.

Hollow *Sphagnum* species are able to establish on bare peat surfaces (Li and Vitt, 1995; Campeau and Rochefort, 1996), as widely varying water availability is common to both disturbed and undisturbed hollows. Hummock species are at a disadvantage, as they are not able to utilize their pore structure and physiology which is adapted to desiccation avoidance, not tolerance (Wagner and Titus, 1984; Vitt, 2000). In *Sphagnum* "sheep" the moisture regime is hypothesized to be similar enough to pre-disturbance conditions to allow for the re-establishment of desiccation avoiding hummock species, with the accompanying faster growth. In contrast, the hollows are expected to have a more severe wetting and drying cycle during the growing season, such that *Sphagnum* growth is less extensive, slower, and in competition with true mosses and lichens. By sampling the surface *VWC* and pore-water pressure of both hummocks and hollows in a post-fire peatland, soil moisture thresholds can be related back to the spatial distribution of newly regenerated *Sphagnum* mosses. Price and Whitehead (2001) used a similar methodology in their investigation of the patterns of *Sphagnum* recolonization in a cutover peatland in Québec and found thresholds of *Sphagnum* colonization similar to laboratory thresholds for *Sphagnum* desiccation.

1.3. Wildfire in the Context of Climate and Land Use Change

Future stress on forested peatlands will occur on multiple fronts, with climate change impacts being compounded by future wildfire scenarios. GCM simulations of regional climate in the boreal plains ecozone (Price *et al.,* 2011) show minor increases in precipitation and vapour pressure under the most likely of IPCC scenarios, with such changes occurring alongside an increase in temperature primarily in the autumn through spring seasons. Such future climates will likely prolong the snow-free season and thus extend the duration and magnitude of evaporation and transpiration with only a marginal increase in precipitation. The most probable outcome of such a future climate is longer, drier summers with more frequent and larger *WT* drawdown extending later into the autumn. Whether this climate scenario will result in a dramatic decline in peatland carbon storage, as predicted by Ise *et al.*, (2008), is uncertain but it is likely a function of the strength of the negative feedbacks contained within peatlands (Belyea, 2010). Concurrent to this climate-induced stress on peatlands will be an increase in burned area in Canada between 10-300%, depending on the emissions scenario and model used. Given that forested peatlands in Canada have a fire return interval of between 90-110 years (Turetsky *et al.*, 2004), an increase in the burned area could result in as little as a 40 year fire return interval for forested peatlands, which would allow for little to no time for the recovery of the carbon lost during the fire during inter-fire peat accumulation period (Wieder *et al.,* 2009). Whether forested peatlands have sufficient fuel to sustain a severe fire at such time intervals, however, is unknown. This study aims to provide an understanding of wildfire impacts on forested peatlands under the current climate, with observations and modelling made in a manner that will assist in the future forecasting of fire and climate change impacts.

Alongside climate forcings on peatlands, energy development in the boreal plains of Canada will likely have a significant role in future peatland disturbance. Currently, linear disturbances such as roads and railways cover only 0.5% of the land surface area of the boreal plains of Alberta, but occur at a density of 0.3 km km^{-2} (ABMI, 2011). These disturbances are not re-routed to avoid peatlands, and can be assumed to occur in equal frequency across upland forests and forested peatlands. Road construction across forested peatlands impounds water on the upslope side and creates a WT drawdown on the other, which dramatically alters tree growth (Lieffers *et al.,* 1987). The alteration of vegetation structure also presumably changes wildfire fuel loading and potential fire severity and ecosystem impacts from fire. Moreover, there exists currently 2.0 km km^{-2} of lower-impact linear disturbances such as seismic cutlines, pipelines, and powerline rights-of-way that do not directly impact the flow of water, but instead clear trees from the peatland surface. Such alterations likely have profound impacts the energy and water balance of peatlands and the post-fire vegetation recovery trajectory. The study of wildfire disturbance in a natural setting will provide a baseline set of observations on the process ecohydrology of natural disturbance, with which the above-mentioned anthropogenic impacts can be compared.

1.4. Study Objectives and Experimental Design

This study aims to provide a process ecohydrology understanding of the impacts of wildfire on forested ombrotrophic peatlands in the boreal plains of Alberta, Canada. Particular consideration will be given to *Sphagnum* mosses, as they are not only the dominant ground vegetation cover and carbon fixation agent, but also are the origin of the vast majority of the peat substrate. As the primary driver of the water balance, the energy relations (while not a full energy balance) will be examined in the context of the energy available for evaporation before and after wildfire. The water balance will be computed and examined to determine both the net change in the water balance after fire, and also the contribution and sensitivity of the various processes contributing to the water balance. The water balance is ultimately just a lower boundary condition for the peatland largely dictating the *WT* and thickness of the unsaturated zone. Thus, the physical properties of peat water retention will be examined in the context of water retention in the unsaturated zone, as well as in the control of the rate of *WT* response to drying and wetting. Lastly, hydrological and ecological processes will be integrated in the examination of *Sphagnum* stress and post-fire environment as it pertains to the recolonization of the peat surface by *Sphagnum* after fire.

This study focuses on a single peatland complex in northern Alberta, approximately 60 km north of the town of Slave Lake. A detailed physical and ecological description of the peatland are available in chapters 2-5. We focused our studies of the impacts of wildfire on a 8 ha wildfire scar within the peatland. The fire was of human origin, caused by a short in a powerline in September 2006. This site was referred to as "BC06" or "burned crow" after the suspicion by firefighters that the short was caused by a bird landing on the powerline. We chose an adjacent area 800 m away of similar stand structure and subject to the same stand-replacing wildfire as BC06 (prior to the one of study) in approximately 1935. This site was referred to as "BC35". Similar instrumentation was installed at both sites, and the placement and frequency of measurement at the two sites was kept as similar as possible. Initial instrumentation was installed in the autumn of 2007, and an initial field season was undertaken in the summer of 2008. By the spring of 2009, the complete set of instrumentation at both sites was installed, and observations were completed by August 2010. The majority of the data collected was from automated instrumentation though important data were made manually. Laboratory analysis was largely limited to the measurements of peat water retention properties, the subject of Chapter 4.

1.5. References

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CHAPTER 2: EFFECT OF WILDFIRE ON FORESTED PEATLAND ALBEDO AND RADIATION EXCHANGE

2.1. Abstract

Forested boreal peatlands represent a precipitation-dependent ecosystem that is prone to wildfire disturbance. Solar radiation exchange in forested peatlands is modified by the growth of an heterogeneous, open-crown tree canopy, as well as its likely disturbance from wildfire. Radiation exchange at the peat surface is important in peatlands, as evaporation from the peat surface is the dominant pathway of water loss in the system. We examined short and longwave radiation exchange in two forested ombrotrophic peatlands of central Alberta, Canada: one with (> 75 years since wildfire; burned) and another without a living spruce canopy (1-4 years since wildfire; unburned) between the autumn of 2007 and 2010. Winter albedo was greater in the recently burned peatland, owing to the loss of the living tree canopy. Accordingly, the unburned peatland showed larger positive net radiation fluxes to the surface. During the summer, rapid regrowth of shrubs reduced the impact of the dark charred peat on surface albedo. While above-canopy net radiation declined slightly in the summer after wildfire, incoming shortwave radiation at the peat surface was greatly enhanced due to wildfire. The associated decrease in longwave losses did not offset these shortwave radiation gains. Radiation transmission through the canopy became more spatially uniform after wildfire, with the burned peatland shifting to a radiation regime resembling the most open patches

of a treed, unburned peatland. After wildfire, this increase in incident radiation substantially increases the amount of energy available for evaporation at the peat surface.

2.2. Introduction

Forested peatlands of the boreal plain in western Canada represent both a predominantly precipitation-dependent wetland and an open-crown forest of *Picea mariana* and *Larix laricina*. Forested peatland carbon stocks are dominated by the surface *Sphagnum* moss community and underlying peat, with the vascular plants accounting for only a small portion of the overall carbon stock in the ecosystem (Wieder *et al.,* 2009). Forested peatlands are subject to wildfire disturbance on average between 80 and 110 years, only marginally longer wildfire return interval compared to uplands (Turetsky *et al.,* 2004). Forested peatlands at the southern extent of the boreal forest are predicted to be at greatest risk of future drought and severe wildfire disturbance (Flannigan *et al.,* 2005).

Aspects of radiation exchange such as albedo can act as a strong negative radiative forcing after disturbance of conifer stands in the boreal forest (Betts, 2000; Lohila *et al.,* 2010). After wildfire in the boreal forest, increases in albedo can offset the positive forcing of direct carbon emissions from organic soil and vegetation combustion and indirect effects such as black carbon deposition (Randerson *et al.,* 2006). A substantial source of the albedo-induced negative forcing is from the transition to a deciduous stand from conifer after wildfire (Randerson *et al.,* 2006). However, in forested ombrotrophic peatlands, deciduous succession is not possible due to the acidic and water-logged conditions in the ecosystem (Benscoter and Vitt, 2008).

Wildfire also extensively alters short and longwave radiation exchange in conifer canopies through the removal of the needles and fine branches. In winter and early spring, Liu et al. (2005) found the increase in albedo from the loss of the conifer canopy reduced the annual net radiation flux in a newly burned Alaskan black spruce forest by 18%. Following wildfire in a spruce-lichen woodland, Rouse and Kershaw (1971) observed a small decrease in summer net radiation (Q^*) due to increased longwave losses. Similarly, Amiro et al. (2006) found that wildfire caused an increase in both summer and winter albedo in a boreal forest chronosequence of conifer and deciduous stands.

Since the short and longwave radiation regime of the ground layer is a major contributor to the evaporation demand in peatlands (Price *et al.,* 1998), changes in the radiation regime following wildfire will also likely contribute to changes in peatland water storage. While this availability of water is important for regenerating Sphagnum mosses (Price and Whitehead, 2001), radiation inputs at the peat surface also influence the temperature regime (Kettridge and Baird, 2010) of the peatland and hence play a role in peat decomposition and methane emissions (e.g. Moore and Dalva, 1993). However, contemporary models of peatland temperature dynamics (e.g. Kettridge and Baird, 2010) currently do not account for the presence of a tree canopy in a burned or unburned state when calculating radiation exchange in peatlands and the associated wildfire disturbance. Moreover, given that a changing boreal climate will most likely include a warmer and drier climate (Bonan, 2008) that may foster tree growth in peatlands (c.f. MacDonald and Yin, 1999) and an accelerated fire cycle (Wotton *et al.,* 2010), there is a need to quantify the impact of trees on radiation exchange in peatlands and the associated wildfire disturbance.

Here we used modelling of radiation attenuation and exchange within and below the canopy to determine how the radiation balance (Q^*, K^*, L^*) of the peat surface changes after wildfire. We hypothesized that wildfire leads to an increase in Q^* at the peat surface after wildfire during the summer due to a decrease in canopy light attenuation and that the alteration of the winter-time radiation regime would be similar to an upland conifer forest.

2.3. Site Description

Field studies were conducted in a forested ombrotrophic peatland (55.87° N, 115.10° W) partially disturbed by wildfire in 2006. The remainder of the peatland was last disturbed by wildfire in approximately 1935 (hereafter referred to as BC35). The 2006 burn site (hereafter referred to as BC06) is part of a 44 ha wildfire from September 2006, killing all trees in the wildfire scar and burning over 99% of the peat area within the wildfire scar. Mean depth of peat consumption at BC06 was estimated to be 7.5 cm using the adventitious root method of Kasischke et al. (2008). The 1935 burn site (BC35) is located 350 m from the edge of the 2006 burn.

These peatlands are typical of forested boreal, ombrotrophic peatlands of the continental high boreal wetland region (NWWG, 1986). The local abundance of peatlands on the landscape is 27% (Tarnocai *et al.,* 2000). The long-term average (1971-2002) annual precipitation and temperature for Slave Lake is 502 mm and 1.7ºC, respectively (Environment Canada, 2000) with an average annual open-water evaporation of 585 mm (Bothe and Abraham, 1993). Snowfall is a relatively small proportion of precipitation (29%), even though the snowpack is present for an average of 144 days per year (Environment Canada, 2000). The winter of 2010 equalled the lowest recorded April 1st snowpack (0 mm), and 6th lowest March 1st (38 mm) in a 35 year record near Slave Lake, AB (Alberta Environment, 2010).

Vegetation at both sites is dominated by *Sphagnum* mosses in a pattern of high areas 30-40 cm in elevation and spanning 1 m^2 or more (hummocks) and low features (hollows). Hummocks are largely composed of *Sphagnum fuscum* with a shrub canopy of *Ledum groenlandicum* that can reach 30-50 cm in height. *Rubus chamaemorus*, *Vaccinium vitis-idaea*, and *V. oxycoccus* grow immediately adjacent to the moss surface, and do not reach more than 5-10 cm in height. Hollows are a mix of *S. angustifolium*, *Cladonia spp.* lichens, *Pleurozium schreberi* feathermosses, and bare peat, with generally a lower density of shrubs. Trees at BC35 are exclusively *Picea mariana* with a mean height of 2.3 m at both sites and a basal area of 11.0 and 11.2 m^2 ha^{-1} at BC06 and BC35, respectively. The stem density is 19 700 and 16 050 stems ha^{-1} at BC06 and BC35, respectively.

Mean peat depth is 1.6 and 1.7 m at BC06 and BC35, respectively. The post-wildfire peatland environment is typified by largely unburned *Sphagnum* sheep (*sensu* D. Vitt), which are composed of *Sphagnum fuscum* hummocks, interspersed with more deeply burned hollows with a thick char layer. BC06 contains 62% hummocks by area, while 72% of BC35 is classified as hummocks.

2.4. Methods

At each site, above-canopy net radiation (Q^*_a) was measured at 10 m height by a NR-Lite net radiometer (Kipp and Zonen: Delft, Neatherlands) starting in October 2007. Above-canopy incoming ($K{\downarrow}_a$) and reflected ($K{\uparrow}_a$) shortwave radiation was measured by two separate LP02 pyranometers (Hukseflux: Delft, Neatherlands) also at a height of 10 m starting June 2008. $K{\downarrow}_a$ was not measured at the unburned (BC35) site, but was assumed to be identical as the burned (BC06) site 800 m away. Net longwave radiation (L^*_a) was calculated as the residual of Q^*_a and net shortwave radiation (K^*_a). Radiation data were recorded on a CR1000 (Campbell Scientific: Logan, Utah) datalogger scanning every 30 s and averaging every 20 min.

In order to capture an unbiased sample of surface processes in a patterned landscape of hummocks and hollows, an orthogonal 40 m by 40 m grid was created at each site. Grid points at a 5 m spacing where surveyed from a random starting location, with the axes corresponding to the cardinal directions.

Below-canopy albedo (α_b) was measured using two Eppley precision spectral pyranometers (Eppley Labs: Newport, RI) and recorded using a voltmeter accurate to 0.1 mV. Measurements were taken at an elevation of 30 cm above the peat surface, and include the shrub canopy. A roving CNR1 radiometer (Kipp and Zonen: Delft, Neatherlands) was deployed to measure below-canopy net radiation (Q^*_b) at 1 m height above the peat surface for up to two weeks at a time during the summers of 2010.

Hemispherical canopy photos were taken at 40 cm above the peat surface using a Nikon D60 fitted with a Sunex (Carlsbad, CA) 185º super-fisheye lens during the summer of 2010. Canopy photos were analyzed with the Gap Light Analyzer (GLA) software (Frazer *et al.,* 2000) which has been shown to be a suitable substitute for measured solar radiation in both open and closed canopy conifer forests (Hardy *et al.,* 2004). Photos were analyzed for tree canopy transmittance (τ_a) as an average value across May-September.

An LP-80 ceptometer (Decagon: Pullman, WA, USA) was used to measure shrub canopy transmittance (τ_b) in July 2009 when maximum leaf-out conditions occurred. Shrub transmittance measurements primarily capture the *Ledum* canopy, and do not account for other shrubs that grow immediately adjacent to the moss surface. For each of the grid points at each site, the total shortwave radiation availability at the surface (K^*_s) was calculated at the product of light attenuation and surface albedo (α_s):

$$K_s = \tau_a \tau_b \left(1 - \alpha_s\right) \tag{1}$$

because moss α_s was not able to be measured in the absence of a shrub canopy, α_s was set to 0.05, corresponding to measurements of freshly burned peat in the summer of 2011 at burn scars within 10 km of the study site, while α_s for lightly burned hummocks (< 2 cm depth of burn) and unburned peat were set to 0.12 (Berglund and Mace, 1972; Kellner, 2001). When multiplied by a radiative efficiency (the ratio of $K*_b$ to $Q*_b$), the flux of $K*_s$ can be considered the amount of radiation available at the peat surface for evaporation.

2.5. Results

2.5.1. Albedo

During the snow season, noon above-canopy albedo (α_a) at the BC35 site averaged 0.33 and 0.37 in the 2008-2009 and 2009-2010 winters, respectively (Figure 2.1). At the recently burned BC06, α_a was significantly higher between 0.57-0.66 (t_{366} = 33.8; $P < 0.001$; Figure 2.1). At BC06, the reduction in snowfall in the 2009-2010 winter resulted in a winter α_a of 0.10 less than the previous winter (t_{190} = 8.4; $P < 0.001$) and a reduction in the duration of the snow season from 155 days to 107 days. At BC35, a small increase in the mean winter α_a of 0.05 was observed in 2009-2010 (t_{88} = 6.5; $P < 0.001$), despite the thinner snowpack.

The month of August was used as a measure of peak summer conditions (owing to the continuity and fully-leaf-out conditions), α_a was significantly higher at BC06 compared to BC35 in 2008 (t_{58} = 34.3; $P < 0.001$) and 2009 (t_{58} = 4.5; $P < 0.001$), though the difference was larger in 2008 compared to 2009. Similarly, α_a varied between years at the same site: August α_a was higher in 2009 compared to 2008 at both BC06 (t_{58} = 58.4; $P < 0.001$) and BC35 (t_{58} = 28.5; $P < 0.001$). In August 2010, α_a was not significantly different between sites nor different from 2009 values at the same site.

Below-canopy (α_b) measured after leaf-out was 0.165 and 0.205 at BC06 and BC35, respectively, with a mean albedo difference of 0.035±0.012 when averaged across four sampling dates (Figure 2.2; $t_{548} = 5.69$; $P < 0.001$). Hummocks prior to leaf-out showed a slightly greater albedo at both sites, ($t_{160} = 1.69$; $P = 0.085$), though the difference was not significant at the 95% CI. Albedo prior to leaf-out measured in late May at all microforms showed a significantly lower surface albedo at BC06 (0.09) compared to BC35 (0.20), a mean difference of 0.11±0.02 ($t_{160} = 10.35$; $P < 0.001$). Compared to α_a, α_b was consistently greater, except for the pre-leaf out measurement at BC06. The difference between α_a and α_b was also site dependent: below-canopy albedo exceeded above-canopy measurements by approximately 0.02 at BC06 after leaf-out, while the difference was 0.05-0.10 at BC06.

2.5.2. Net Radiation

Trends in above-canopy net radiation (Q^*_a) are largely similar to the differences in albedo. On an annual scale, Q^*_a was 10-12% greater at BC35 compared to BC06. Both sites experienced the onset of negative monthly Q^*_a in November. Positive Q^*_a commenced in February at BC35, but was delayed by one month at the burned site (Figure 2.3). This delay in the onset of consecutive daily Q^*_a fluxes appears to represent the source of annual variability between sites. February had the largest contrast in Q^*_a between sites, followed by March and January. May and September are notable in that Q^*_a is larger at BC06 compared to BC35 in those months only. While mean daily Q^*_a in a particular month was never significantly different between sites, the difference in daily Q^*_a was significantly different than zero in every month, with only May and September where Q^*_a at BC06 consistently exceeded that at BC35.

Winter diel patterns in net radiation at BC35 show an enhanced positive Q^*_a flux to the surface during the day (Figure 2.4), while Q^*_a remained negative at BC06. Longwave losses exceed shortwave gains during the day at BC06, but not BC35. Only BC35 shows any period of consistent positive Q^*_a at any part of the day during the course of the winter. Mean radiative efficiency (the ratio of $Q^*_a/K\downarrow_a$) in summer was 0.53 and 0.54 at BC35 and BC06, respectively. A 10% decrease in Q^*_a during the day was observed at BC06 (Figure 2.4). While K^*_a is roughly similar between sites, longwave losses during the day are 12% larger at BC06 compared to BC35. While outgoing longwave radiation ($L\uparrow_a$) was not directly measured, we can assume that incoming longwave radiation is the same at both BC06 and BC35, owing to their close proximity (only 800 m apart). Using the Stefan-Boltzmann law and a surface emissivity of 0.98, the difference in longwave radiation corresponds to an average daytime surface temperature increase from 20°C to 32°C. Such increases in surface radiative temperature are supported by sub-canopy measurements of longwave radiation (Figure 2.6). However, the exponential nature of the Stefan-Boltzmann law implies that small areas of very warm surface could be disproportionately influential in determining longwave fluxes above the canopy ($L\uparrow_a$).

2.5.3. Below-canopy Processes

Mean canopy transmittance (τ_a) is greater at BC06 than BC35 (t_{160} = 14.1; $P <$ 0.001) There is very little overlap in the distribution of τ_a between BC06 and BC35; the 5th percentile of τ_a at BC06 (0.86) corresponds to the 80th percentile in the unburned canopy of BC06 (Figure 2.5). Variance in τ_a decreased after wildfire, with the variance in

τ_a decreasing by 68% at BC06 compared to BC35 ($F_{80,80} = 11.41$; $P < 0.001$). Radiation interception within the shrub canopy was greater in the unburned BC35 ($\tau_b = 0.35$) site compared to BC06 ($\tau_b = 0.66$). Differences in the variance of τ_b between BC35 and BC06 was not significant ($F_{80,80} = 1.02$, $P = 0.92$) and there was no correlation between τ_a and τ_b ($R^2 = 0.025$, $P = 0.155$) at either site.

Measurements of below-canopy shortwave radiation are consistent with canopy hemispherical photos. The $K{\downarrow}_b$:$K{\downarrow}_a$ ratio was greater than one at the surface in BC06 and at a very open site (hummock) at BC35 (Table 2.1); shortwave radiation may be reflected by trees immediately to the north of the surface radiometers, but do not otherwise provide shade. $K{\downarrow}_b$:$K{\downarrow}_a$ at the densely treed hollow at BC35 was 34%, similar to the value of 40% calculated via canopy photography. The understory at the BC35 hollows shows a highly variable shortwave radiation regime, consistent with shadowing from a patchy tree canopy. Observations of Q^*_b in the BC35 hollow understory were negative during the day 17% of the time (900-1600h LST) while Q^*_a was positive. In comparison, no such decoupling of Q^*_a to Q^*_b was observed in the relatively more open BC35 (Figure 2.6). Autocorrelation of Q^*_b was significant at a 95% CI at BC06 and the open hummock in BC35 to a lag between 3.6 and 4.3 hours, in the heavily treed BC35 hollow, autocorrelation is only significant to a lag of 1.3 hours, suggesting a more variable shortwave radiation regime.

Longwave losses also decreased by over half at the heavily treed BC35 hollow compared to the more open hummock at BC35 and both locations in BC06. Between-site differences in longwave losses were greatest during the day, with the hollows at BC06 having the largest longwave losses. Despite the decreases in longwave losses, the ratio Q^*_b:$K\downarrow_a$ in the heavily treed area of BC35 is 0.2, only 28% of the values at BC06. Overall, the surface radiation regime of the open gap at BC35 is nearly identical to that of BC06, despite the large differences in the composition of the moss surface.

2.5.4. Radiation Availability at the Peat Surface

Total shortwave radiation to the peat surface (K^*_s) was calculated as the product of the tree canopy, shrub, and surface albedo values (Figure 2.6). Mean K^*_s at BC35 was lower (0.32), while K^*_s was higher at BC06 (0.56), indicating a dramatic increase in shortwave radiation and therefore radiation available for evaporation at the peat surface. While variance in K^*_s did not vary between sites, the 1st quartile of K^*_s at BC06 corresponds to the 3rd quartile at BC35, indicating a large shift in the amount of shortwave energy input. Areas of BC06 with little shrub cover receive over 10 times more incoming shortwave radiation compared to the densely treed areas of BC35.

2.6. Discussion

2.6.1. Role of Trees Before and After Wildfire

The impact of the removal of the conifer cover is most apparent in the differences in winter albedo (Figure 2.1). Conifer trees have been widely shown to shed snow and maintain a low albedo throughout a boreal winter (Leonard and Eschner, 1968; Arain *et al.,* 2003), however our observations of winter albedo in burned peatlands are lower compared to unforested peatlands. For example, Lafleur *et al*. (1997) reported a spring snow-on albedo of 0.7 in an unforested fen in central Saskatchewan, Canada while

Berglund and Mace (1972) observed a winter albedo of 0.81 in an unforested bog in Minnesota, USA. Winter albedo in boreal closed-canopy conifer forests has been shown to range from 0.20-0.25 (Arain *et al.,* 2003; Amiro *et al.,* 2006), with winter albedo values as low as 0.10 for a heavily treed peatland (Berglund and Mace, 1972). The difference in winter albedo between unburned and burned sites in this study (0.25-0.35) is smaller than the 0.55 difference observed in a chronosequence of boreal conifer and mixwood stands in Amiro *et al*. (2006). This can likely be attributed to the patchy and open crown nature of the conifer overstory in peatlands compared to the relatively more closed-canopy upland conifer forest. The role of trees in summer albedo is muted by the similarity between a closed-canopy conifer forest albedo (*c*. 0.08; Arain *et al.,* 2003) and the burned or unburned peatland surface (0.05 and 0.10). While the tree canopy is sparse, the distributed and upright nature of the trees means they play a disproportionately large role in determining above-canopy albedo, despite only occupying a small portion of the total surface area of the peatland.

When considering the short and longwave radiation balance, the presence of trees in forested peatlands studied here are fundamentally different from unforested peatlands that have traditionally been the focus of most peatland science. For example, median τ_a is 0.75, but can be as little as 0.44 in densely treed patches. Canopy transmittance at BC35 is still much larger than closed-canopy conifer forests where τ_a is below 0.3 (Bisbee *et al.,* 2001; Hardy *et al.,* 2004). The impact of wildfire appears to shift the entire peatland to a shortwave-dominated radiation regime similar to that of only the most open and treeless patches of an unburned peatland. Similar to shifts in winter albedo, the magnitude of change in canopy transmittance in a forested peatland appears smaller than the 90-100% increase in canopy transmittance observed in closed-canopy forests (Burles and Boon, 2011; Leach and Moore, 2011).

Observations of understory longwave radiation exchange during the summer in conifer forests are limited. Baldocchi *et al*. (1997) observed an 80-90% decline in Q^* between the top of a closed-canopy black spruce stand and the forest floor. Late winter snowmelt measurements of longwave radiation in closed-canopy forests show a dramatic shift after wildfire disturbance, where longwave radiation losses are greatly reduced or a net positive flux to the surface (Burles and Boon, 2011). However, the enhanced longwave losses we observed at burned surfaces do not exceed the increased shortwave radiation to the surface as a result of the removal of the tree canopy. Our results also suggest that the impact also appears to converge longwave emissions from the peatland surface and the top of the canopy as our above-canopy and below-canopy observations of L^* closely track one another at BC06, while the fluxes diverge during both the day and night at BC35. Consequently, the peat surface is more exposed to longwave losses at night after wildfire. The increase in nighttime longwave losses following wildfire could also lead to earlier frosts and frost heaving at the peat surface, which has been shown to suppress *Sphagnum* colonization in peatlands (Quinty and Rochefort, 2000). However, this potential for increased nighttime cooling in burned peatlands may enhance dewfall, which has been shown to constitute a significant water input to mosses in temperate grasslands (Csintalan *et al.,* 2000). Moreover, increased longwave radiation losses at night enhances distillation, which is the condensation of warm, humid air from the shallow soil horizons on the cool moss surface (Carleton and Dunham, 2003). The distillation process was found to contribute significantly to the water inputs of feathermosses in late summer by Careton and Dunham (2003), though the contribution of distillation in higher-moisture systems such as peatlands is unclear.

2.6.2. Shrub Impacts on Radiation Exchange

Shrub attenuation of shortwave radiation was on average greater than the attenuation from the conifer canopy at all three sites. Moreover, the range and standard deviation in shrub LAI at all sites exceeds the variation in canopy transmittance, thus enhancing spatial variability in an already patchy landscape. While patchy trees on the landscape provide transient shading highly dependent on solar azimuth (Giesbrecht and Woo, 2000), shrubs act as a filter to radiation following Beer's law, and are likely less dependent on zenith or azimuth. While trees can act upon the peat surface by decreasing $K\downarrow$ sufficiently to render Q^* negative at mid-day (Figure 2.6), shrubs appear less likely to have such binary impacts on solar radiation (Sonnentag *et al.,* 2007), though continuous measurements of radiation exchange underneath shrubs were not undertaken in this study.

A shrub albedo of approximately 0.10 exceeded the peat surface albedo in hollows at BC06, while the opposite was true in the BC06 hummocks and throughout BC35. Thus, shrubs appear to have a moderating effect on α_b by decreasing α_b before fire while increasing α_b after fire. The relatively rapid regrowth of shrubs compared to the spruce canopy also shows the important role shrubs play in modifying the radiation balance of these forested bogs. While radiation observations were not made until one year after wildfire in October 2007, ericaceous shrubs such as *Ledum groenlandicum* were observed in abundance at BC06 by that time. By 2008-2010, LAI was greater than 1.0 in some locations, thus moderating the albedo impact of the char surface. The similarity of trends in α_a between sites in the same year suggests that the annual variability in shrub growth may mask any impacts of moss regrowth on α_a in the first four years following wildfire. However, the peatland ecosystem is limited in its abundance of shrubs compared to an upland due to the ombrotrophic nature of peatlands where succession of *Populus*, *Salix*, or *Betula* trees to further increase surface albedo (i.e. Randerson *et al.,* 2006) is not possible in this ecosystem.

2.6.3. Radiation Balance of the Peat Surface

The impact of wildfire on the surface radiation balance of forested peatlands is a different response compared to upland forests. The presence of *Sphagnum* sheep (*sensu* D. Vitt), *Sphagnum fuscum* hummocks that are only burned to less than 2 cm or merely singed, moderates what would otherwise be a greatly reduced surface albedo. While wildfire appears to be a process that creates a more spatially uniform regime of canopy shortwave transmission, wildfire also increases the contrast between hummock and hollow surface albedo. Before wildfire, vegetation such as *Cladina spp*. lichens or feathermosses in hollows increased albedo beyond that of *S. fuscum* hummocks. However, the greater depth of combustion in those same hollows (Benscoter and Wieder, 2003; Benscoter *et al.,* 2011) leads to greater char formation that in turn reduces albedo.

2.7. Conclusions

We used micrometeorological observations and measurements of vegetation structure in forested boreal peatlands of western Canada to characterize the impact of wildfire on peatland radiation balance. The impact of wildfire on albedo is only apparent in the presence of snow and before leaf-out, where we observed a marked increase in albedo in recently burned sites. The regrowth of a shrub canopy and the presence of char-free *Sphagnum* "sheep" ameliorate changes in albedo between burned and unburned. Increased longwave radiation losses caused a roughly 10% decline in net radiation for peatlands within four years since wildfire, as observed from the above- canopy.

While the above-canopy summertime radiation balance changed only slightly, the impacts on energy availability at the peat surface were dramatic. The complete combustion of the open conifer canopy led to a higher and more uniform transmittance of solar radiation through the canopy. Radiation at the peat surface increased more than ten fold in some areas, while in unburned areas within tree gaps, the surface radiation balance differs little from burned areas. The relatively rapid regrowth of shrubs within one year serves to further moderate changes in albedo and to increase shortwave radiation attenuation to the peat surface.

Though the observations of post-wildfire radiation exchange presented here largely concur with upland forests, some different ecological features of peatlands, such as the presence of *Sphagnum* hummocks and a lack of deciduous trees and large shrubs, confer a contrasting radiative response. The trends and processes the findings identified in this study will assist future efforts to parameterize episodic wildfire disturbance in peatlands into earth system models.

2.8. Acknowledgements

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Table 2.1. Below-canopy radiation fluxes at burned and unburned hummocks and hollows, expressed as a ratio to the above-canopy $K\downarrow_a$ flux; n is the percentile of canopy transmittance for the location compared to other photographs at the same site.

Site	Microform	τ_a	n	$K\downarrow_b$	$K\uparrow_b$	K^*_b	L^*_b	Q^*_b
BC35	Hollow	0.40	0.01	0.34	-0.06	0.29	-0.08	0.20
BC35	Hummock	0.93	0.98	1.02	-0.14	0.89	-0.17	0.71
BC06	Hollow	0.96	0.61	1.03	-0.13	0.90	-0.22	0.68
BC06	Hummock	0.94	0.53	1.03	-0.14	0.89	-0.22	0.67

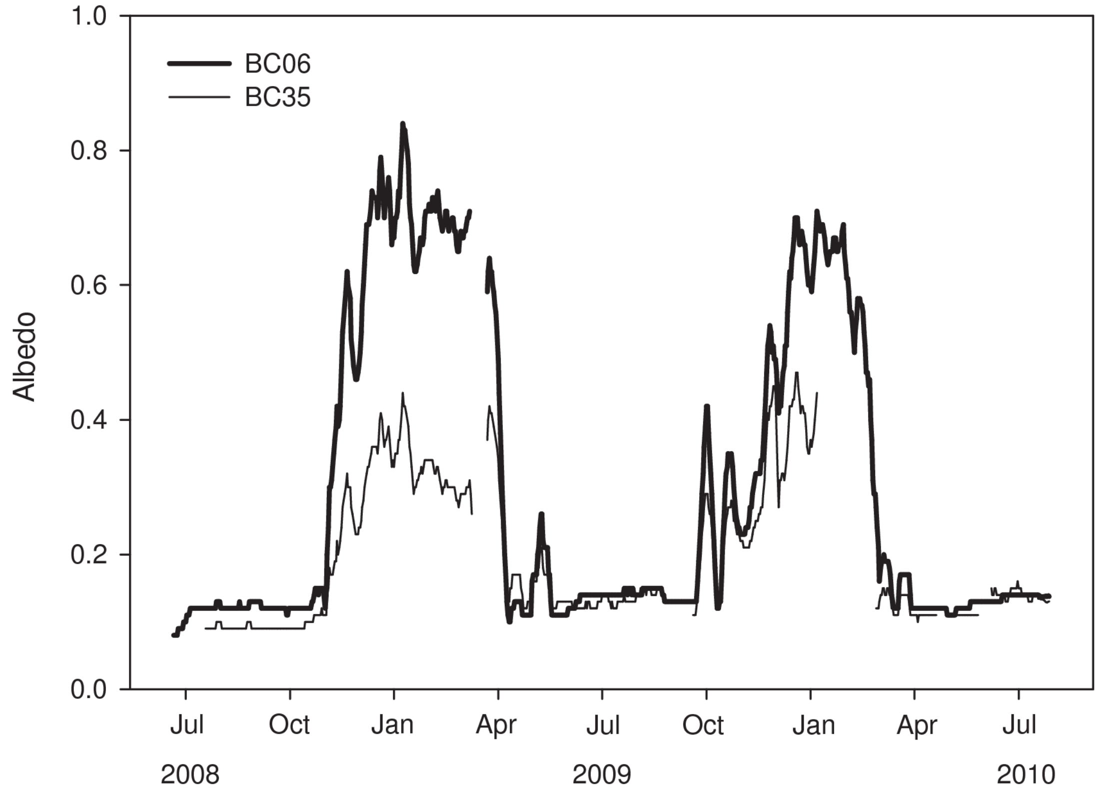

Figure 2.1. Above-canopy albedo (α_a) plotted as a three-day running average from July 2008 to September 2010.

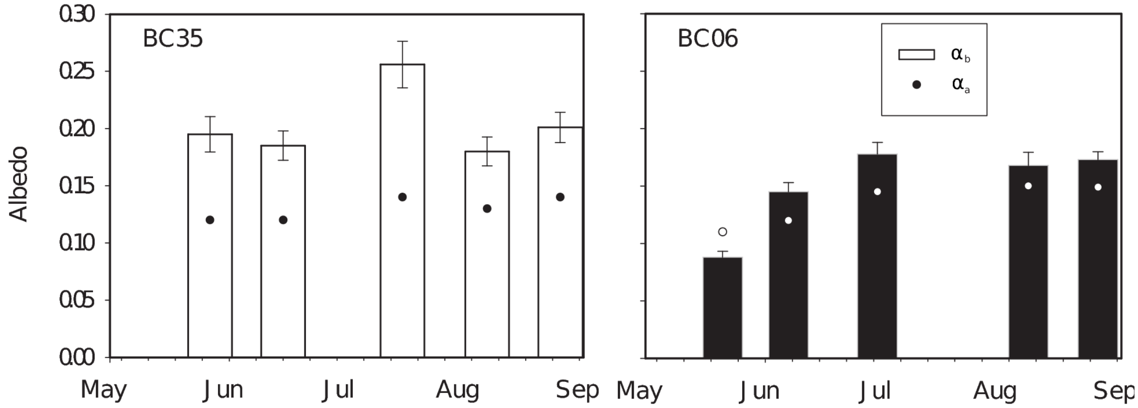

Figure 2.2. Comparison of above-canopy (circles) versus below-canopy albedo (bars) measured in 2009. The first measurements at each site represents pre-leaf out conditions.

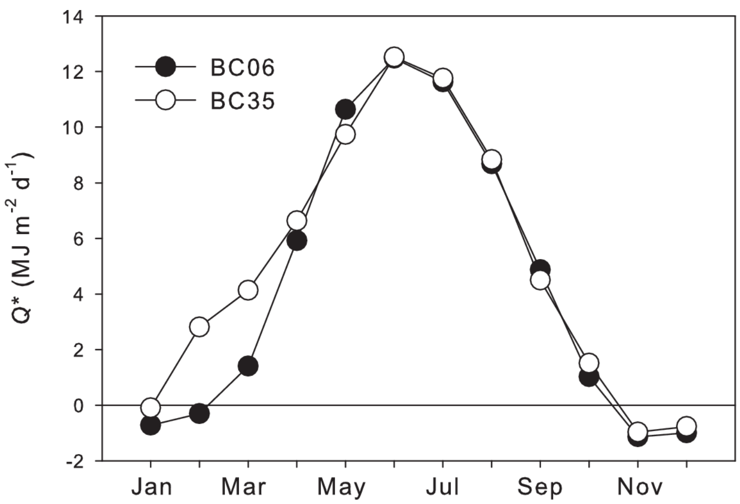

Figure 2.3. Daily above-canopy net radiation (Q^*_a) by month. Values are averaged from October 2007-August 2010.

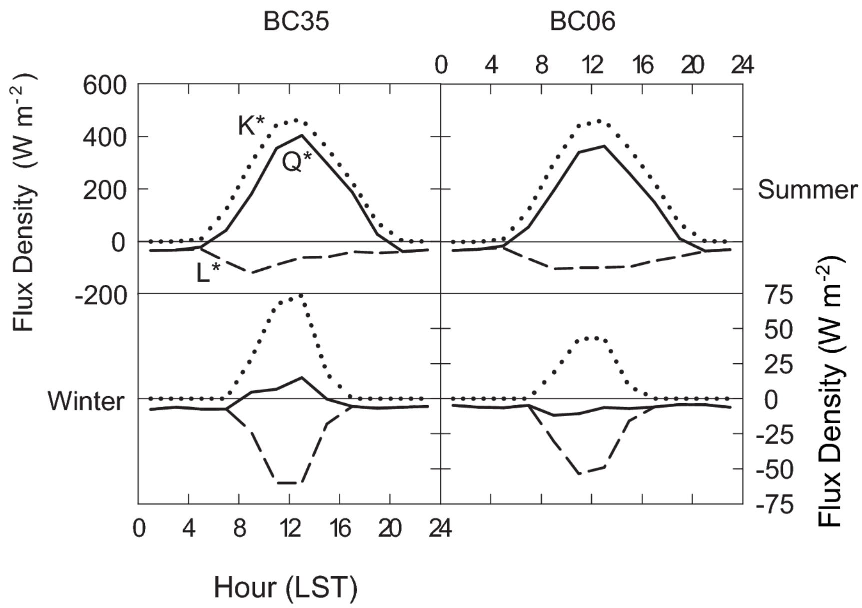

Figure 2.4. Diel trends in above-canopy radiation exchange, as a ten-day average over a two hour interval in January and August 2010.

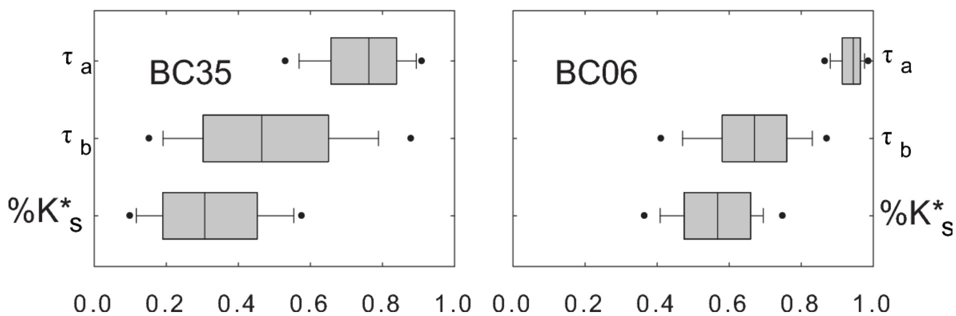

Figure 2.5. Proportion of incoming shortwave radiation transmission through the tree (τ_a) and shrub (τ_b) canopy. $\%K^*_s$ is the proportion of $K\downarrow_a$ that is absorbed by the peat surface. The boxplots represent the distribution of individual measurements taken at 81 points at each site.

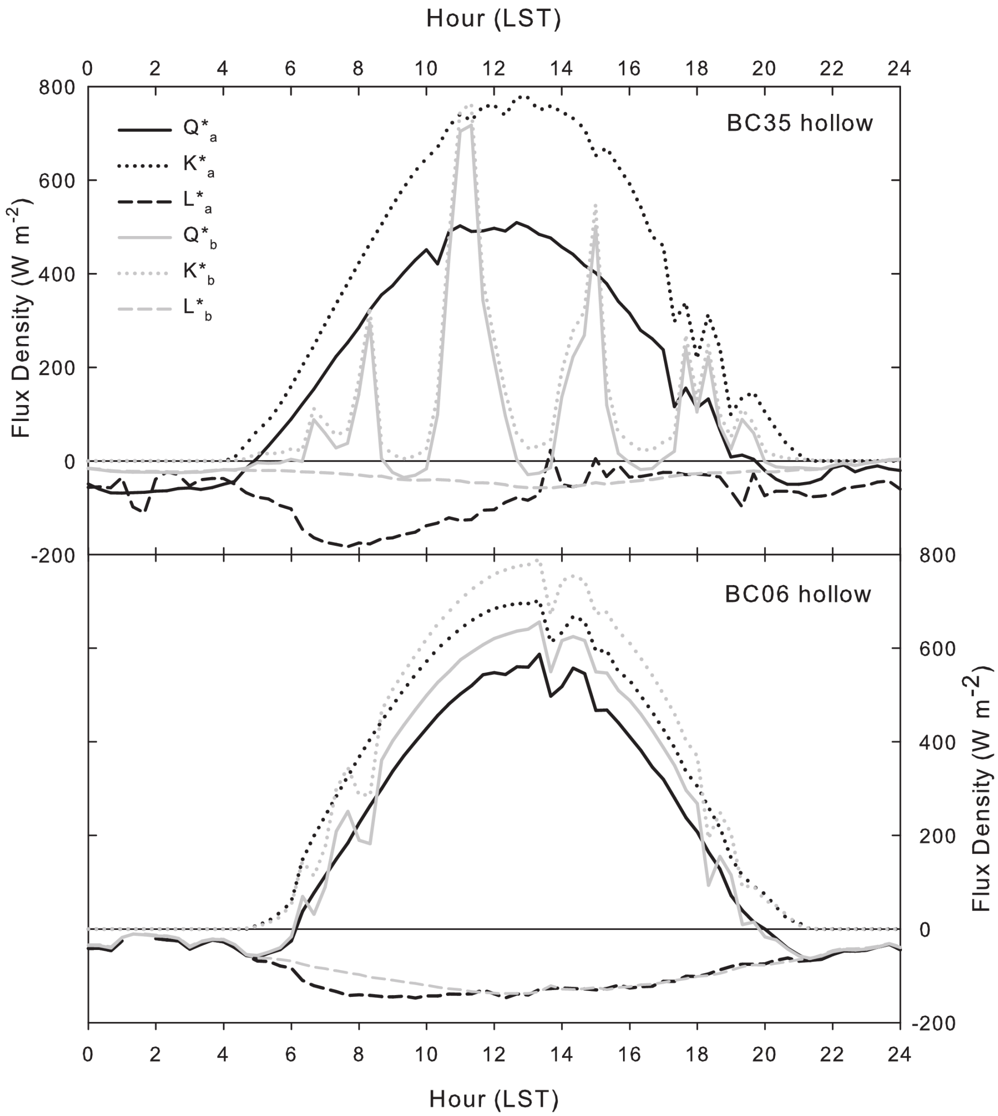

Figure 2.6. Comparison of above and below-canopy radiation at BC35 (upper panel) and BC06 (lower panel) over a 24 h period. The top and bottom panels show observations from Julian days 182 and 196, 2010.

CHAPTER 3: CONTRASTING THE WATER BALANCE OF A BURNED AND UNBURNED FORESTED PEATLAND

3.1. Abstract

The impact of wildfire disturbance on the water balance of a forested ombrotrophic peatland in the boreal plain of western Canada was observed over a three year period. Complete combustion of foliage and fine branches dramatically increased shortwave radiation inputs to the peat surface while halting all tree transpiration. End-of-winter snowpack increased 7-25% after fire due to decreased ablation from the tree canopy. Shrub regrowth was rapid, and shading by the shrub canopy in the burned site approached that of the unburned site within three years after fire. Surface resistance to evaporation did not change significantly after fire, but aerodynamic roughness and aerodynamic resistance decreased slightly. Water loss at both burned and unburned peatlands is largely driven by surface evaporative losses. The role of trees in modifying the evapotranspiration regime of peatlands is discussed in the context of post-wildfire water balance. Increased evaporation after fire marginally exceeded the sum of pre-fire transpiration and interception, leading to a net loss of 20-40 mm of water during the growing season in the burned peatland. The majority of increased evaporative demand after wildfire was drawn from the saturated zone, as unsaturated storage decreased only slightly. Specific storage decreased after wildfire, but was balanced by larger water table declines, resulting in no change in the peat surface elevation regime after fire. Moreover, specific yield increased after fire more than specific storage decreased, leading to a more

variable water table and larger summertime water table decline after wildfire. Large rain events in excess of 10 mm showed a distinctly greater proportional water table response, such that water table modeling was only successful if additional water table rise including the capillary fringe was incorporated.

3.2. Introduction

The hydrological response to wildfire disturbance in forested catchments has been widely shown to increase streamflow (*e.g.* Berndt, 1971; Inbar *et al*., 1998; Neary *et al*., 2004) largely owing to the complete loss of transpiration (Amiro, 2001). Moreover, numerous field investigations and modelling studies (*e.g.* Rouse, 1976; Amiro, 2001; Bond-Lamnerty *et al*., 2009) have shown a reduction in overall evapotranspiration immediately following wildfire in upland boreal forest stands. The decrease in transpiration after wildfire may cause an increase in water storage rather than increased runoff, as is the case in forested hillslopes affected by fire (Inbar *et al*., 1998). While a majority of studies have focused on streamflow as the main hydrological response to wildfire disturbance, over 20% of the Canadian boreal forest is composed of peatlands (Tarnocai *et al*., 2000), where streamflow or runoff is only a minor component of the water balance and water storage in the saturated and unsaturated zone represents a better indicator of hydrological response (Van Seters and Price, 2001).

While the hydrological impacts of direct anthropogenic disturbances such as harvesting are well known (Van Seters and Price, 2001), stand-replacing wildfire is the largest disturbance by area in North American peatlands (Turetsky *et al*., 2004). Rowe and Scotter (1973) suggested that fire in forested peatlands in boreal western Canada increases the height of the water table due to the loss of tree transpiration, but had no evidence to prove their hypothesis.

The water balance and associated soil moisture in peatlands after wildfire disturbance is an important control over post-fire moss recolonization, particularly of *Sphagnum* mosses (Benscoter, 2006) which require adequately high soil moisture to recolonize (Price and Whitehead, 2001). While *Sphagnum* mosses can fully recolonize within 25 years after fire (Benscoter and Vitt, 2008), the return of the conifer canopy can take upwards of 50 years (Wieder *et al*., 2009). Conversely to *Sphagnum* mosses, the establishment and growth of black spruce in forested peatlands is slowed by a higher water table and soil moisture (LeBarron, 1945).

In forested and afforested peatlands, previous studies have shown trees and shrubs to play a major role in peatland hydrology by simultaneously lowering surface moisture via transpiration and interception (van Seters and Price, 2001) and providing shade that reduces surface evaporative demand on the moss surface (Chapter 2). In order to quantify the net impact of trees on the peatland, both the water-losing (transpiration, interception) and water-conserving (shading of the peat surface) impacts of trees need to be considered. In the case of closed-canopy forested peatlands that have been drained to enhance forestry yields (*i.e.* Juntras, 2006), it is apparent from the increase in water table position after harvesting that the net effect of trees was to maintain a lowered water table, largely via transpiration. Natural, open-canopy forested peatlands, have shown a decrease in water table position after wildfire (Simard *et al*., 2007) suggesting that trees may have a water-conserving effect in undisturbed systems. Similarly, Rothwell (1982) observed a 40 cm water table decline after timber harvesting in a forested wetland in Alberta, though the water table was lower at the harvested site compared to unharvested adjacent wetlands only in drier years. The contribution of specific processes to the net loss of water after wildfire and the ways in which trees impact the water balance of peatlands in the drier and more continental climate of the western Canadian boreal forest remains unknown.

This study aims to: (i) document the impact of wildfire on water storage in a western Canadian ombrotrophic peatland, and (ii) examine which specific hydrological process is responsible for the majority of change following wildfire.

3.3. Site Description

Observations were made within a peatland complex (55.8º N, 115.1º W) 70 km north of Slave Lake, Alberta, Canada. The region is classified as continental mid-Boreal forest (NWWG, 1986), typified by long, dry winters, with summers being dominated by convective precipitation. Mean annual precipitation at Slave Lake (1971-2000) was 502 mm, 146 mm of which falls as snow (Environment Canada, 2000). Mean annual temperature was 1.6 ºC, with a mean July temperature of 15.6 ºC (Environment Canada, 2000). Peatlands in the local area are typically underlain by glacial clay deposits. These low-gradient peatlands are classified as "flat bogs" and lack surface runoff channels (National Wetlands Working Group, 1988).

The peatland complex is composed primarily of ombrotrophic forested bog that last burned *c.* 1935 (BC35) and an 8 ha portion of the peatland that burned in September 2006 (BC06). Mean combustion at BC06 as measured after the fire by the aventitious root method (Kasischke *et al*., 2008), was 7.5 cm. Hydrometeorological towers were installed in the middle of the BC06 burn scar and an adjacent unburned stand (BC35) was installed 800 m away in October 2007.

Vegetation at both sites was dominated by *Sphagnum* moss hummocks 30-40 cm high and approximate 1 m^2 in area, as well as low hollows. Hummocks occupy 72% of the landscape at BC35, and 62% at BC06. Peat depth averaged 1.6 m and 1.7 m at BC06 and BC35, respectively. Hummocks were largely composed of *Sphagnum fuscum* with a shrub canopy dominated by *Ledum groenlandicum* that can reach up to 30 cm in height. Other shrubs and forbs, including *Rubus chamaemorus, Maianthemum trifolia, Vaccinium oxycoccus,* and *V. vitus-idaea* occupy less than 25% of the hummock biomass. Hollows are a mix of *Sphagnum angustifolium*, lichens, and bare peat. After fire, the hummocks sustained little combustion, with most hummocks experiencing less than 2 cm of combustion. Hollows experienced extensive combustion, resulting in a blackened char and ash layer and no remnants of living moss. The hollows are primarily bare peat with less than 20% *Polytrichum strictum* cover. *Vaccinium sp.* shrubs were absent from the burned site.

Tree cover at both sites is composed entirely of an open canopy of black spruce (*Picea mariana*) with a basal area of 11 m^2 ha^{-1} and an average height of 2.3 m and all stems at BC06 are fire-killed *P. mariana* snags. Stem density was 19 700 and 16 000 stems ha^{-1} at BC06 and BC35, respectively. Canopy cover (f_{tree}) was 0.27 at BC35, and was calculated as only the area occupied by tree boles (0.001) at BC06.

3.4. Methods

3.4.1. Water Balance

The water balance of low-gradient, ombrotrophic peatlands can be modelled entirely as a two-dimensional process (*e.g.* Ingram, 1982), where the change in storage, ΔS is controlled by precipitation (P), precipitation interception (I), evaporation (E), transpiration (T), groundwater flow (Q):

$$\Delta S = P - I - E - T \pm Q \pm \Delta\theta \quad (3.1)$$

Using snowmelt as the time for model initiation, water deficits at snowmelt (S_0) must be accounted for, since snowmelt does not always replenish water deficits in western Canada (Lawson and Dalrymple, 1996). In addition to calculated values of ΔS, the water table was modeled (WT_m) on a daily timestep:

$$WT_m = \frac{(\Delta S)}{(S_y + bS_s)} \pm S_0 \quad (3.2)$$

where S_y is the depth-averaged specific yield, b is the thickness of the saturated peat, and S_s is the specific storage. S_y was calculated from laboratory tests from peat at the same sites (Chapter 4). S_s was calculated from the relationship between WT and peat surface elevation, as recorded using 16 mm diameter rebar inserted into the peat down to the clay substrate underlying the peat (Price and Schlotzhaur, 1999). During rain events, the water table rise may be in excess of that predicted by (S_y + bS_s) alone, as the addition of water to the capillary fringe causes a rapid rise in the water table with little to no volume of water added (Gillham, 1984). The water table response from a single rain event can be modelled on a daily scale:

$$\Delta WT = \frac{(P - ET)}{(Sy + bS_s)} + z_{cf} \tag{3.3}$$

where z_{cf} is the water table rise due to the capillary fringe. This value can be esimated from pore size characteristics or water retention tests (Gillham, 1984). Based on the laboratory tests presented in Chapter 4, we estimated the capillary fringe to be 50 mm and 25 mm as BC06 and BC35, respectively.

The water table level was continuously logged beside the tower at each site in a hollow using either a Dataflow Systems (Christchurch, NZ) Odyssey capacitance water table logger or an Ott (Kempten, Germany) PLS pressure transducer in a 5 cm diameter PVC well installed to a depth of 1.5 m. Unsaturated zone storage was monitored using arrays of Campbell Scientific (Logan, UT, USA) CS616 moisture probes inserted at 5, 15, and 30 cm depths in a hummock and a hollow at each site. Further details of the calibration of the probes are described in Chapter 5.

3.4.2. Groundwater Flow

Surface runoff channels were not observed in the vicinity of the peatland, so all lateral losses were assumed to be as a result of saturated flow. Steady state groundwater losses from the bog dome (mm d^{-1}) was calculated using the groundwater mound model of Ingram (1982) as modified for variable saturated hydraulic conductivity with depth by Morris and Waddington (2011):

$$Q=\sum_{i=1}^{n}\frac{HK_i\,\Delta z}{L^2} \tag{3.4}$$

where H is the height of water table above the base of the peat (m), K_i is the saturated hydraulic conductivity (m d^{-1}), Δz is the thickness of the peat layer (m), and L is the width of the peatland (m). The ratio of peat thickness to peatland radius (z/L) is equivalent to the mean hydraulic gradient. Hydraulic conductivity of the peat was determined using the pump test method of Hvorslev (1951) with peizometers at depths of 0.5, 0.7, 1.0, and 1.2 m below the hollow surface. While the peatland complex east of Utikuma Lake is over 12 km in width, the local ombrotrophic portion of the complex which both sides reside was estimated from satellite imagery at 1800 m, and the smaller value of L was utilized in the calculations. As the water table declined during the course of the summer, the values of H and Δz were adjusted accordingly for the uppermost layer measured at 0.5 m and representative of 0-0.6 m depths.

3.4.3. Precipitation and Interception

In addition to a logging rain gauge at the BC06 tower, throughfall in trees (I_{tree}) was monitored by logging tipping bucket rain gauges, which were placed under the dripline of a tree at BC35 and at the same distance from a burned tree bole of the same diameter at BC06. Interception by shrubs way from trees (I_{open}) was measured by a series of manual rain gauges deployed away from trees in BC35. Shrub interception was not directly measured at BC06, but was scaled based on the relative difference in shrub LAI. Landscape-scale interception was calculated as:

$$I=\left(\left(1-f_{tree}\right)I_{open}\right)+\left(f_{tree}\,I_{open}\right) \tag{3.5}$$

where f_{tree} is the canopy cover, and I_{open} and I_{tree} are the interception away from and underneath the canopy of a conifer tree, respectively.

3.4.4 Evaporation

The Penman-Monteith evaporation model (Monteith, 1965) was used to model peat surface evaporation (E). Net radiation (Q^*) was measured above the canopy with Kipp and Zonen (Delft, Neatherlands) NR Lite net radiometers at 10 m elevation. Ground heat flux (Q_g) was calculated using the combination gradient and calorimetric method (Liebethal *et al*., 2005), as traditional heat flux plates often underestimate ground heat flux in peat soils (Halliwell and Rouse, 2006). The ground heat flux was calculated as the gradient across the 10 cm and 20 cm depth, along with the change in energy storage in the 0-10 cm layer as calculated by thermocouple (20 gauge type T, Omega Engineering: Samford, CT, USA) measurements at 2 and 5 cm. Ground heat flux was measured at one hummock and one hollow at each of the sites.

3.4.5. Radiation

In order to calculate Q^* at the evaporating peat surface, incoming shortwave radiation at the top of the canopy ($K\downarrow$) was modified by observed light transmittance through the tree canopy (τ_a) obtained via hemispherical photography with a Nikon D80 digital camera and a Sunex Superfisheye lens. Photos were taken on a low tripod at 30 cm elevation, above any shrubs. Canopy photos were anlayzed via the Gap Light Analyzer software (Frazer *et al*., 1998) at 81 points and averaged across each site. Light transmittance through the shrub canopy (τ_b) was measured using a LP-80 ceptometer (Decagon: Pullman, WA, USA). A simplified version of the LAI model of Norman and

Jarvis (1974) was used to calculate LAI from light transmittance beneath the shrubs. Shrub transmittance was combined with values obtained from hemispherical photos taken at the same location in order to calculate a surface-level Q* at 81 locations at each site:

$$Q = K\downarrow \tau_a \tau_b \mathrm{RE} \tag{3.6}$$

where RE is the summertime understory radiation efficiency of 0.65 for both sites (Chapter 2).

3.4.6. Ground heat flux

The volumetric heat capacity of the peat was calculated using an average bulk density value of 30 kg m^{-3} for hummocks and 60 kg m^{-3} for hollows, along with volumetric water content measurements from a Campbell Scientific CS 616 moisture probe inserted at an angle from 0-20 cm. Outputs of the CS 616 probes were transformed into volumetric water contents using the mixing model of Kellner and Lundin (2001) with a modification for the differing probe circuitry of the CS 616 using the calibration of Hansson and Lundin (2006).

3.4.7. Resistances

Aerodynamic resistance (r_a) of the trees was calculated as a function of wind speed at 10 m height using a wind profile (Allen *et al*., 1998) of three RM Young (Traverse City, MI, USA) model 03101 anemometers. Displacement height (d) was modelled using the simple relation from Abtew (1989):

$$d = h_c f_{tree} \tag{3.7}$$

where h_c is canopy height in metres, and f_{tree} is the tree canopy cover (a value different than canopy openness as measured by a densiometer, Jennings et al., 1999).

Aerodynamic resistance from shrubs was not measured via the wind profile technique, but detailed micrometeorological studies have shown that a dense canopy of short shrubs and grasses in treed peatlands confer an r_a of approximately 60 s m^{-1} (Spieksma *et al.*, 1997), which was added to the aerodynamic resistance of BC35. We used a shrub r_a of 35 s m^{-1} at BC06, based on a shrub LAI being half that of BC35.

Surface evaporation was measured at 10 locations at BC06 and 11 locations at BC35, each 17 cm in diameter using closed-chamber measurements with a PP Systems EGM-4 infrared gas analyzer. The model of Griend and Owe (1994) was used to compute surface resistance (r_s) from surface evaporation. Peat temperature, manually measured at 2 cm depth, was used as the temperature at which to base partial pressure of water vapour in the peat pore space for the resistance model of Griend and Owe (1994). Where 2 cm measurements were missing, a multiple linear regression model based on air temperature and soil temperature at 10 cm was used ($R^2 = 0.77$; $F_{2,434} = 747$; $SE_y = 2.6$°C). A total of 196 surface resistance observations were made between both sites in 2008 and 2009. The geometric mean of hummock and hollow measurements at each site was weighted by the proportion of hummocks (f_{hum}) and hollows (f_{hol}) on the landscape to derive a site-averaged r_s value:

$$r_s = \left(f_{hum} r_{hum}\right) + \left(f_{hol} r_{hol}\right) \tag{3.8}$$

Shrubs, primarily *Ledum groenlandicum*, were included in the chamber measurements during measurements of r_s. Supplemental measurements of stomatal resistance from the shrub canopy (r_c) were taken separately from surface resistance in July 2010 on 20 specimens using a Delta-T Devices (Cambridge, UK) AP4 porometer at each site.

3.4.8. Transpiration

Transpiration (T) was modelled as an independent process from surface evaporation. Sapflow observations were made on three trees: one representing the lower 30%, one the middle 40%, and one the upper 30% of the within-site distribution of basal area at BC35 using thermal dissipation probes (Dynamax Inc, Houston, Texas, USA). Measurements of 31 trees revealed an average sapwood to basal area ratio (A_s:A_t) of 0.44. Daily T (mm d^{-1}) was then calculated as the weighted average of the daily sums of sapflow in each of the three trees. Sapflow was only measured in the 2010 field season. For the 2008 and 2009 field season, daily transpiration was modelled using the relationship between transpiration and daily average vapour pressure deficit normalized to the length of the day (D_z; Oren *et al.*, 1996) derived from the 2010 measurements. Transpiration was assumed to begin within 10 days of the last spring snowfall or hard frost, corresponding to Julian Day 143 and 120 in 2009 and 2010, respectively. The Penman-Monteith model was run on a 20 minute time-step and summed to daily values. Transpiration was modelled on a daily timestep. The water balance model was run during the 2008-2010 field seasons.

3.5. Results

3.5.1. Water Table and Specific Yield

Specific yield as measured by lab analysis of peat cores (Chapter 4) at the site did not differ between sites in the hummocks at 0.50 (t_{38} = 0.51, P = 0.615). There was a small decrease in S_y at the hollows from 0.26 at BC35 to 0.21 at BC06, though the difference is not significant (t_{36} = 0.613, P = 0.526). When the observed S_y values for hummocks and hollows are multiplied by f_{hum} and f_{hol}, the average S_y for BC35 and BC06 are 0.42 and 0.39, respectively.

Peat depth (b) averaged 1.5 m in the hollows at both sites. Both BC35 and BC06 experienced similar surface elevation change of 5 cm over the course of the 2009 summer. However, the peat elevation change at BC35 occurred during a water table decline of 35 cm, while at BC06 the same elevation change occurred over a 50 cm water table decline. The resultant value of bS_s calculated for each site was 0.14 at BC35 and 0.10 at BC06, since the same elevation change required a larger change in water table at BC06.

In the spring of 2008, the water table at BC06 was approximately 8 cm lower than that at BC35. By Julian Day (JD) 205, *WT* declined by 11 cm at BC06, while at BC35 the *WT* decline was larger at 14 cm (Figure 3.1). During the spring of 2009, the *WT* position at BC06, as measured to a hollow surface, was 2 cm lower than BC35. By JD 245, *WT* position at BC06 was 17 cm lower than BC35. In contrast, a water table difference of 18 cm was observed in the low-snowfall spring of 2010, with the difference narrowing to 9 cm by the end of the season. One significant feature of the spring of 2010 was that the water table at the burned site was below the depth of seasonally frozen peat as of JD 125, while the water table was frozen at the same time at BC35.

The magnitude of *WT* change differed dramatically between years, though the response was similar between sites given their water balance. In 2009, initially high snowfall was followed by rain, resulting in water tables at or near the hollow surface.

The remainder of the summer of 2009 was marked by small and infrequent rainfall, resulting in a large *WT* decline of 500 and 370 mm at BC06 and BC35, respectively (Figure 3.1). In contrast, the summer of 2010 was marked by low *WT* and storage deficits, even after snowmelt (Table 3.1; Figure 3.1). Larger and more frequent rainfall events occurred in 2010, however, resulting in a much smaller *WT* decline. In particular, the large rain events of greater than 10 mm had a distinctly large water table response, disproportionate to smaller rain events on the order of 2-5 mm.

3.5.2. Snow and Throughfall

Snow water equivalent (SWE) in the spring of 2008 was significantly greater at BC06, at 119 mm SWE compared to 89 mm SWE at BC35 (t_{291} = 8.5, P < 0.001). Mean end of winter snowpack in 2009 measured 96 and 89 mm SWE at BC06 and BC35, respectively, a mean difference of 7%. However, given the equally large variance in snowpack between the sites (F_{44} = 1.06; P = 0.85), this difference was not significant at a 95% CI (t_{88} = 1.8; P = 0.075). The winter of 2010 was equal to the driest winter on record, with a complete loss of snow cover by March 1st (Environment Canada, 2010). Snow was a large contributor to the annual water balance of 2008 and 2009, equal to as much as 71% of rainfall received at BC35 and BC06 during the study period from the beginning of May to the end of August.

Cumulative rainfall interception underneath the dripline of black spruce trees in BC35 was 45% and 65% in 2009 and 2010, respectively. Manual rain gauges showed an additional 10% interception by the shrub canopy at BC35, with an estimated 5% shrub interception at BC06. With f_{tree} equal to 0.27, I was calculated to be 20% and 26% of P

in 2009 and 2010 at BC35. This amount of interception is equal to 9% of E during the study season in 2009 and 24% of E in 2010. While trees only covered 27% of the surface area of the peatland (f_{tree}), they were responsible for 69% of the interception, with I_{tree} being over six times that of I_{shrub}. In contrast, I in BC06 was equal to only 2-6% of E, split roughly equally between the burned tree stems and the shrub canopy.

3.5.3. Aerodynamic Resistance

Displacement height (d) for the logarithmic wind profile was calculated to be 0.63 and 0.0026 m at BC35 and BC06. Roughness length for momentum was 0.22 m at BC35, but decreased to 0.14 m at BC06. Median aerodynamic resistance to vapour flux (r_a) was 30 s m^{-1} BC35, similar to 32 s m^{-1} measured at BC06 (Figure 3.2). At the median 10 m elevation wind speed of 2.1 m s^{-1}, r_a was 44 and 45 s m^{-1} at BC35 and BC06, respectively.

3.5.4. Net Radiation and Ground Heat Flux

Daily canopy-top net radiation was marginally greater at BC35 in all months except May. Canopy photography showed an average of 74% canopy transmittance at the unburned site over the months of May-August. After wildfire and combustion of fine fuels in the canopy, transmittance increased to 94%. The impact of shading by the living tree canopy at BC35 was quantified by comparing Q^* measured at the ground level at 20 minute intervals during the study period. This difference in shading by trees corresponds to a average loss of 27 W m^{-2} to the surface of BC35 and upwards of 100 W m^{-2} near noon. Shrub LAI, equal to 0.84 and 0.44 at BC35 and BC06, resulted in a further shading of the peat surface by 50% and 33%, respectively. Average net radiation at BC35 was

37% of incoming shortwave radiation at the top of the canopy. At BC06, this proportion increased to 62%. Q_g in hummocks at BC35 averaged 1.5% of Q^*, while hummocks at BC06 averaged 1.5% of Q^*. Ground heat flux at the hollow at BC35 was on average 6.8% of Q^*, but was only 3.2% of Q^* at BC06. When the flux is weighted by f_{hum} and f_{hol}, Q_{g} of both sites averaged 3% of daily Q^*.

3.5.5. Transpiration

Daily transpiration ranged between 0.2 and 1.2 mm d^{-1}, with an average of 0.7 mm d^{-1}. Cumulative transpiration over study periods in 2008-2010 was 54-75 mm (Table 1). Transpiration was highly correlated with both the daily sum of photosynthetically active radiation (PAR; R^2 = 0.80; F_{36} = 141; P < 0.001) and vapour pressure deficit (Figure 3.3; D_z; R^2 = 0.81; F_{36} = 156; P < 0.001), although PAR and D_z were themselves highly correlated (R^2 = 0.78; F_{36} = 124; P < 0.01) during the observation period. Mean D_z was not significantly different at a 95% CI between years, with an average between 0.9 and 1.1 kPa. Modelled transpiration values as a function of D_z from Gower *et al*. (2005) were well correlated (r = 0.88) to sapflow measurements, but measurements of daily T exceeded modelled values by an average of 25%.

3.5.6. Surface Resistance

Mean stomatal resistance of *Ledum groenlandium* (r_c) in BC35 was lower at 359 s m^{-1}, compared to 542 s m^{-1} in the burned site (t_{38} = 3.5; P < 0.001; Figure 3.4). In *Rubus chamaemorus*, r_{c} at both sites 250 s m^{-1}, not significantly different between sites (t_{41} = 0.25; P = 0.81). Surface resistance (including both peat and shrubs) in the hollows (r_{hol}) did not differ between BC35 and BC06 sites at 78 s m^{-1} while r_{hum} increased slightly from

56 to 63 s m^{-1} from BC35 to BC06. Once weighted for the distribution of hummocks and hollows on the landscape via Eq. 5, BC06 had a slightly higher r_s of 69 s m^{-1} compared to 62 s m^{-1} for BC35 (Figure 3.5). If the peat and shrub system is considered in a parallel circuit analogue, then the low stomatal conductance (0.80 and 2.3 mm s^{-1} at BC06 and BC35, respectively) of the shrubs suggests that shrubs only contribute 6 and 15% of the total surface conductance at BC06 and BC35, respectively.

3.5.7. Evaporation

Mean daily surface evapotranspiration in the summers of 2008-2010 at BC35 was 2.2 mm d^{-1}, while BC06 had a larger average daily surface evapotranspiration of 3.4 mm d^{-1} (t_{316} = 48.0, $P < 0.001$). Evaporation over the duration of the study periods in 2008-2010 (May-August) was 80-160 mm greater at BC06 compared to BC35 (Table 3.1; Figure 3.6). If the shading from trees was to be taken away and a daily average of 27 W m^{-2} (2.3 MJ m^{-2} d^{-1}) was added to the surface Q^* (see Chapter 2), surface E at BC35 would increase by an average of 0.72 mm d^{-1}, or approximately 60-85 mm over the study season.

In each year the increase in E at BC06 was only partially offset by the lack of T and lower I at BC06, with the net effect being a greater ΔS deficit at the end of each study season at BC06. In 2009 and 2010 the sum of T plus I at BC35 was equal to 70% of the total increase in E at BC35. In 2010, this margin was smaller, with T plus I equal to 90% of enhanced E at BC06. In other words, if either T or I increased at BC35 by between 10 and 30%, the transpiration and interception effects of trees would exceed enhanced E post-fire, and the unburned environment would be drier than that after fire.

3.5.8. Groundwater Flow

K_{sat} varied from 51 m d^{-1} for loose peat at 50 cm depth to 0.002 m d^{-1} for peat at 1.2 m depth, only 20 cm above the base of the peatland itself. Given the logarithmic decline in K_{sat}, the upper 0.6 m was calculated to be responsible for over 99% of the saturated flow. A WT decline of 20 cm below the hollow surface resulted in a 50% decline in already small Q values, while a 50 cm WT drop lead to a 90% decline in Q. Calculated hydraulic gradients from the Ingram model were 0.0015, while measured gradients averaged 0.0044 and 0.0020 at BC35 and BC06, respectively. Mean daily Q calculated using equation 3 was less than 0.1 mm d^{-1}, and thus far below the errors in the other fluxes measured in this study. Using the larger observed hydraulic gradients, this resulted in values of Q still smaller than the error of the larger fluxes.

3.5.9. Storage

Overall, unsaturated storage losses were a minor component, occupying only ~3% of total storage losses, as the majority of the water storage lost to evaporation came from the saturated zone. Water table decline during the summer was well-correlated with ΔS (Figure 3.7), with a mean ratio of ΔWT:ΔS at the end of each study season (equivalent to depth-averaged $S_y + bS_s$) of 0.65 at BC35 in each year 2008-2010. At BC06 the same ratio ΔWT:ΔS was more variable, being as little as 0.48 in the dry summer of 2009, and as large as 0.71 in the summer of 2008 up to JD 211. Optimal water balance model performance was found using single values of $S_y + bS_s$ equal to 0.65 and 0.50 at BC35 and BC06, respectively (Figure 3.8). Model accuracy declined significantly if S_y values divergent from the optimal were used, though a smaller error was incurred from an increased S_y compared to smaller S_y values.

While ΔS was well-correlated to WT decline during rain-free periods without compensating for the capillary fringe, prediction of the WT was poor after large rain events. Observed WT rise was often 2-3 times what equation 3 would predict without the inclusion of the capillary rise effect, z_{cf}. Excluding the capillary rise results in an error at the end of the 2010 season of 250 and 125 mm and BC06 and BC35, respectively. With less overall rain and fewer large rain events, the impact of the capillary rise on *WT* fluctuation was only equal to 50 and 25 mm at BC06 and BC35, respectively.

3.6. Discussion

3.6.1. Snow

Differences in snowpack between the two sites can largely be attributed to the presence of the unburned conifer canopy at BC35, as conifers such as black spruce have been widely shown to both intercept snowfall (*e.g.* Davis *et al*., 1997; Pomeroy *et al*., 2002) and to enhanced mid-winter ablation via longwave radiation from stems and foliage (Davis *et al*., 1997). A simple snow ablation model (Pomeroy *et al*., 2002) that requires only f_{tree}, predicts that 12% of the end of season snowpack will ablate. This figure is similar to our observations of 21% and 7% ablation in 2008 and 2009. The winter of 2009-2010 differed in that a dry fall without any significant rain prior to freeze-up was followed by a long and cold winter with little snow. Only 22 mm SWE of snow fell between December 2009 and February 2010, and this was followed by a very early melt in the first week of March 2010. As a result, there was widespread freezing of the saturated zone at BC35, while the deeper water table of BC06 limited freezing solely to the unsaturated zone.

Snow was a large proportion of the precipitation input in 2008 and 2009, which is not normal for the region (Environment Canada, 2000). In 2009, the amount of snow was sufficient to lead to a classic post-snowmelt wetting up and rise of the water table to surface, similar to that often observed in maritime peatlands or wetlands with a nival regime (Roulet and Woo, 1986). While snow represented only a major portion of inputs to the water balance in 2009, only when snowmelt was followed by a sufficiently large rainfall of 23 mm over the course of a week was there sufficient water to raise the water table to the surface of hollows at both sites. In 2010, we were unable to measure the snowpack prior to melt, but it is likely that much of the 20 mm SWE snowpack as measured at Slave Lake was ablated away, and any residual water was minor compared to the large storage deficit from the previous fall.

3.6.2. Surface Evaporation

Mean growing season surface evaporation of 3.1 mm d^{-1} at BC06 is similar to the 3-3.5 mm d^{-1} of evapotranspiration measured from a closed canopy black spruce forest in northern Manitoba, Canada, by Arain *et al*. (2003). However, evaporation at BC06 was higher than the evaporation of 2-2.5 mm d^{-1} from an open fen measured by Lafleur *et al*. (1997). Surface evaporation from BC35 averaged only 1.9 mm d^{-1}, and was more similar to measurements of evaporation from the forest floor of closed canopy boreal forests, where *Sphagnum* evaporation is on the order of 1.5 mm d^{-1}, and as little as 0.5 mm d^{-1} in densely shaded areas (Heijmans *et al*., 2004).

Despite the increase in evaporation and greater loss of storage at the burned peatland, the cumulative evapotranspiration did not exceed the mean annual precipitation for the region of 500 mm a^{-1} (Environment Canada, 2000), meaning the peatland will likely not be in year over year water deficits, and any loss in storage is likely to be short term. In contrast to the expected water surplus in the forested peatlands, evapotranspiration in upland aspen stands in the region can be upwards of 530 mm a^{-1} (Devito and Fraser, 2004).

The increase in Q^* at the moss surface after fire was similar to that in upland stands (Leach and Moore, 2010), and was the primary driver of the increase in evaporation after wildfire at BC06. This increase in Q^* appears to be dominated by an increase in incoming shortwave radiation, since non-winter surface albedo does not differ between burned and unburned peatlands after the regeneration of the shrub canopy (Chapter 2).

Ground heat flux was not a significant sink for Q^* at any of the sites and was well within the error involved in modelling surface Q^* in a patchy open canopy forest and the associated use of radiative efficiency to scale $K\downarrow$ to Q^*. Q_g is commonly observed to be a minor sink of Q^* in peatlands (Lafleur, 1997; Petrone, 2004), where soil moisture is high and thermal conductivity is low, and is hence a minor consideration in the Penman-Monteith model.

In the measurement of bulk surface resistance (r_s) in peatlands, it is difficult to distinguish moss and shrub contributions to bulk surface resistance without clipping the shrubs or otherwise altering the system. Values of r_s measured in this study are consistent with the lower range of measurements of r_s in a Swedish ombrotrophic peatland by Kellner (2001), who found r_s to be 160 ± 70 s m^{-1}, including the shrub canopy. The higher shrub and sedge LAI in the peatland observed by Kellner (2001) can largely account for differences between with this study. Moreover, the presence of a generally denser peat owing to the fire disturbance may have increased unsaturated hydraulic conductivity of the peat, thereby likely reducing r_s. This process may be analogous to cutover peatlands, where Petrone *et al*. (2004) measured an increase in evaporation of 5-15% in a cutover peatland compared to an adjacent remediated peatland, despite a 30-35% decrease in surface volumetric water content.

Despite the large sample size of r_s measurements, we were unable to find a statistically significant effect of *WT* depth on r_s, unlike eddy covariance studies of evaporation in peatlands by Kim and Verna (1996) and Lafleur *et al*. (2005). The lack of observation of this effect may be due to due to the fact that this disturbance is not present in forested peatlands, where a generally denser peat increases the unsaturated hydraulic conducitvity (K_{unsat}) and therefore the probability of an increase in r_s based upon evaporative demand rates exceeding K_{unsat}. The lack of statistical significance of this effect could also be in part due to the presence of lichens, feathermosses, and bare peat surfaces, which feature high r_s values regardless of *WT* (Bond-Lamberty *et al.*, 2011). The high inherent variability of r_s measurements, similar to that of K_{sat}, further complicates any direct comparisons. Moreover, there may be a link between high r_s species such as feathermosses and lichens and lower-energy shaded environments (Bisbee *et al.*, 2001), which would enhance spatial heterogeneity and complicate any single deterministic estimates of r_s by point measurement.

3.6.3. Storage and Water Table Modelling

Higher evaporation at BC06 was the main driver of greater storage losses throughout each of the summer study periods. The 50% increase in evaporation at BC06 was partially offset by both a lack of transpiration and decreased precipitation interception, which together were equal to 70% of the evaporation contrast between BC06 and BC35.

While the modelled water table presented here fits well with observations, uncertainty remains in both S_y and to a lesser extent evaporation. Laboratory measurements of S_y were limited to six cores from each site, with only triplicate hollow cores at each site that spanned the entire depth of water table variation. The considerable vertical heterogeneity in peat properties (Chapter 4) is accentuated by the lack of persistence of microforms after fire, creating a complex pattern of alternating hummock and hollow peat of variable composition and S_y. Moreover, S_y can be expected to decrease with depth (Price, 1996) though the rate of S_y decline with depth would conceivably vary between hummocks and hollows, given their contrasting decomposition rates (Turetsky *et al.*, 2008). Encapsulating S_y with a single variable is possible in wetland water balance models (e.g. Price and Schlotzhaur, 1999; Roulet and Woo, 1986), but is unrealistic under a large *WT* range. Some of the decrease in S_y with depth may be coinciding with a decrease in evaporation under a declining water table (Kim and Verna, 1996). The counteracting effects of S_y and evaporation may contribute towards the linearity of water table decline with increasing storage deficits (Waddington *et al*., 2012) which should otherwise not be the case.

3.6.4. Specific Yield and Specific Storage

The water table data presented here use the hollow surface as a common datum between two peatlands. While it provides a common reference point, this convention has many limitations (Waddington *et al.*, 2012), including differences in surface elevation (Price and Schlotzhaur, 1999) and in the relative abundance of hollows on the landscape, which has been shown to change after fire (Benscoter *et al.*, 2005). Moreover, while the relative differences in S_y between hummock and hollow peat are generally understood (Letts *et al.*, 2000), the persistence of high S_y peat within hummocks is poorly understood, such that when using hollows as a broad hydrological datum, the extent to which high S_y peat is found at the same elevation in adjacent hummocks is poorly understood. This has implications for uncertainty when using single values of S_y for modelling purposes.

Overall, $S_y + bS_s$ was smaller at BC06, as shown by the greater WT response to precipitation events (Figure 3.1). The difference in $S_y + bS_s$ resulted in a greater WT response per unit decrease in ΔS. This is particularly important during extended rain-free periods, as the extensive decline of the water table into areas of the peat-profile of increasingly smaller S_y (Boelter, 1964) leads to exponentially greater decline in WT per unit ΔS. This S_y feedback was particularly evident late in the summer of 2009, when water table declines of over 11-12 mm were observed when daily evaporation was less than 4 mm. The peat surface at BC06 was more rigid, as evidenced by smaller values of bS_s. The more rigid peat surface was balanced by a larger water table decline over the summer at BC06, resulting in the same peat elevation change at BC35. Though no pre-fire measurements of S_y are available, post-fire irreversible peat compression, similar to that observed in the cutover peatland environment (Kennedy and Price, 2005), may in part be responsible for the larger S_y values observed here.

3.6.5. Groundwater Flow

Similarly large exponential decreases in K_{sat} with depth have been observed in large, lower-gradient peatlands (Fraser et al., 2001) where no subsurface pipe flow has been observed. Using the model of Ingram (1982), the ratio of the peat thickness to lateral extent is a key control beyond hydraulic conductivity in determining the rate of lateral water losses. In this case, the large later extent coupled with the relatively thin peat present produces relatively low hydraulic gradients compared to a classical dome bogs, where the peat may reach upwards of 8 m thick, but is less than 1 km wide (Morris and Waddington, 2011). Covering approximately 130 km^2, the contiguous nature of the Utikuma Lake peatland complex within which this study is located points to an additional water-conservation advantage for these large peatlands, in that their area to perimeter ratio is minimized. Minimizing the perimeter of peatlands in the boreal plain of western Canada is critical, as transpiration from upland trembling aspen (*Populus tremuloides*) stands induces a steep water table gradient that draws water away from the margin of peatlands. Therefore the configuration and topology of peatlands may play a role in their hydrological resilience, where large and contiguous peatlands may have a distinct advantage by minimizing hydraulic gradients by reducing the relative area adjacent to uplands.

3.6.6. Role of Shrubs

Shrub recovery in peatlands is rapid after wildfire compared to tree growth, with mean shrub LAI having reached 0.44 three years after fire, compared to an LAI of 0.84 in the unburned BC35 site. Mean shrub light transmittance was measured in 2009 at 0.66 and 0.50 at BC06 and BC35, respectively, suggesting that the role of shrubs in shading and reducing Q^* and therefore surface evaporation returns quickly after fire (Chapter 2). Additional shrub growth only marginally increases shading, as self-shading of leaves reduces the shading impact per leaf area. The observed decrease in shrub r_c after fire furthers shrub efficiency in conserving water by reducing transpiration. At BC35, values of r_c are more similar to existing studies. In a Swedish peatland with higher shrub LAI and significant sedge cover, Kellner (2001) estimated shrub stomatal conductance at 350 s m^{-1}, which is very close to the measurements of stomatal conductance given here. Given the net effects of high r_c, the 50% reduction in Q^* and the doubling of r_a, shrubs appear to have a significant role in preserving water in the peatland landscape. In the moderate leaf areas observed both before and after fire, there is still sufficient light for *Sphagnum* growth (Bisbee *et al*., 2001), without the dense shrub growth that crowds out *Sphagnum* growth that is seen in cutover peatlands (Farrick and Price, 2009).

3.6.7. Ecohydrological Impacts of Trees

Transpiration dominates water losses in boreal forest catchments with closed-canopy forests, where soil evaporation is near zero except during snowmelt and after large rain events (Wang, 2008). However, in peatlands low oxygen in the rooting zone (Silins and Rothwell, 1999) suppresses photosynthesis and potentially the transpiration rate of black spruce trees (Dang *et al*., 1991). The approximately 75 year old black spruce stand at BC35 was a minor contributor to the water balance, transpiring only 75 mm a^{-1}, equal to only 34% of *E*. In comparison, closed-canopy spruce stands in the boreal forest can transpire upwards of 325 mm a^{-1}, with only minor surface evaporation (Ewers *et al*., 2005). The proportion of transpiration attributable to trees in BC35 exceeds the 25% value from a drained and partially afforested peatland in Quebec, Canada (Van Seters and Price, 2001).

The consumption of fine needles and branches in BC06 reduced the roughness length by over half from 0.22 m to 0.14. Even at BC35, the roughness length of this open canopy forest is still far less than a closed-canopy forest, where z_0 typically exceeds 1 m (Oke, 1988). In unforested peatlands with a hummock and hollow topography, z_0 can be as little as 0.02 m (Molder and Kellner 2002), showing that trees, even just the burned stems and primary branches, contribute significantly to surface roughness. Given the reduction in surface roughness, aerodynamic resistance (r_a) attributable to the tree canopy was reduced by 20% as a result of wildfire, from 75 s m^{-1} to 60 s m^{-1}. This effect is comparable to the approximately 60 s m^{-1} contributed by shrubs, a value similar in shrubby peatlands both with (Spieksma *et al*., 1997) and without trees (Kellner, 2001). In the absence of trees and shrubs, an aerodynamically smooth *Sphagnum* surface may have a resistance over 300 s m^{-1} (Spieksma *et al*., 1997).

Unlike upland mineral soils where the upper soil horizons are dry and soil surface evaporation is low (Heijmans *et al*., 2004; Bond-Lamberty *et al*., 2011), the abundance of near-surface water in peatlands allows them to evaporate close to equilibrium levels (Petrone, 2004; Brown *et al*., 2010). The 0.7 mm d^{-1} of evaporation saved by the presence of shading of trees at BC35 is similar in magnitude to the 0.7 mm d^{-1}, suggesting that at least in dry periods, the impact of trees on the landscape is water-neutral. However, rainfall interception by the same tree canopy at BC35 averaged 0.4 mm d^{-1} in the wet year 2010, effectively tipping the balance towards a net water losing process when interception is considered. In the drier year 2009, interception by trees was only equal to 0.16 mm d^{-1}, suggesting that trees are a stronger net water losing agent once rain interception is considered. The additional consideration of up to 25 mm SWE of snowpack ablation during with winter contributes up to another 0.1 mm d^{-1} of water loss over an average 200 day snow-free season. However, a few key processes could tip this balance in either direction. An increase in sapwood flux, given the tree density (thus keeping shading and interception constant) would serve to diminish the water-conserving effects of trees in peatlands, as would a shift to smaller, more frequent precipitation events, which favour canopy interception of rainfall (Price *et al*., 1997). An increase in tree density would increase shading only to a point where trees are self-shading, as is the case in closed-canopy forests where the negative feedback of shading is effectively saturated (Gower and Richards, 1990). Once the canopy closes in and *Sphagnum* is outcompeted by feathermosses for light (Bisbee *et al*., 2001), the processes by which *Sphagnum* suppresses the growth of trees (Breeman, 1995) would also decrease, and a transpiration-driven water table drawdown feedback could take hold (Gorham, 1995). However, this transpiration feedback is itself moderated by the aging of the trees, which decreases transpiration per unit sapwood area and per unit leaf area (Ewers *et al*., 2005).

3.7. Conclusions

Forested peatlands of the boreal plain of western Canada experience periodic wildfire disturbance, which significantly alters the physical structure of the peatland by decreasing shading by trees and producing a more aerodynamically smooth surface. Surface resistance decreased only slightly after fire, but was not controlled by depth to water table, suggesting some degree of resilience of evaporative processes at the peat surface after wildfire disturbance. In the post-burn environment, the high stomatal resistance of shrubs coupled with a significant shrub shading effect provides an effective interim water-conserving mechanism in the absence of a tree canopy.

The lack of trees in the burned site also provides interesting insights into the role of trees in unburned peatlands with open-canopy forests. Trees appear to retain water on the landscape in the absence of rainfall, owing to their modest transpiration and ability to shade the peat surface from additional evaporative demand. These water-conserving effects appear to be specific to the intermediate stem densities and low-growth environments observed in the open-canopy forests typical of forested peatlands. During periods of normal or greater precipitation, rainfall interception by trees lessens, and perhaps reverses, the water-conserving effect of trees due to rainfall interception. The effect of upland trees drawing water away from peatlands was not an issue given the location of the study site, though the role of upland-peatland interaction in the resilience of peatlands to wildfire warrants further study.

The decrease in storage and potentially surface moisture availability after fire may impact the viability of *Sphagnum* recolonization due to moisture stresses, thus inducing a positive feedback loop where mosses are outcompeted by more dry-adapted species such as trees and shrubs. An enhanced understanding of vascular plant and moss competition and the implications on the water balance of forested peatlands is essential to predicting water table conditions prior to fire in peatlands, and to estimating the vulnerability of boreal forested peatlands in a changing climate.

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Table 3.1. Water balance component of the burned and unburned peatlands during the study periods in 2008-2010. Values given are relative to zero at the initiation of the water balance, as shown in Figure 3.1. S_0 is not included in the calculations of ΔS shown here.

	2008		2009		2010	
Site	BC35	BC06	BC35	BC06	BC35	BC06
Snow	89	119	89	96	~22	~22
S_0	-62	-46	-22	-15	0	-135
P	182	182	135	135	208	208
I	-37	-14	-27	-14	-57	-22
E	-190	-271	-258	-385	-220	-382
T	-54	0	-76	0	-71	0
ΔS	-99	-103	-226	-264	-140	-196

Table 3.2. Water balance model performance. Mean absolute error (MAE) and mean signed difference (MSD) are measured in mm.

	2009		2010	
Criteria	BC35	BC06	BC35	BC06
MAE	12	22	10	30
MSD	-6	-13	-8	-15

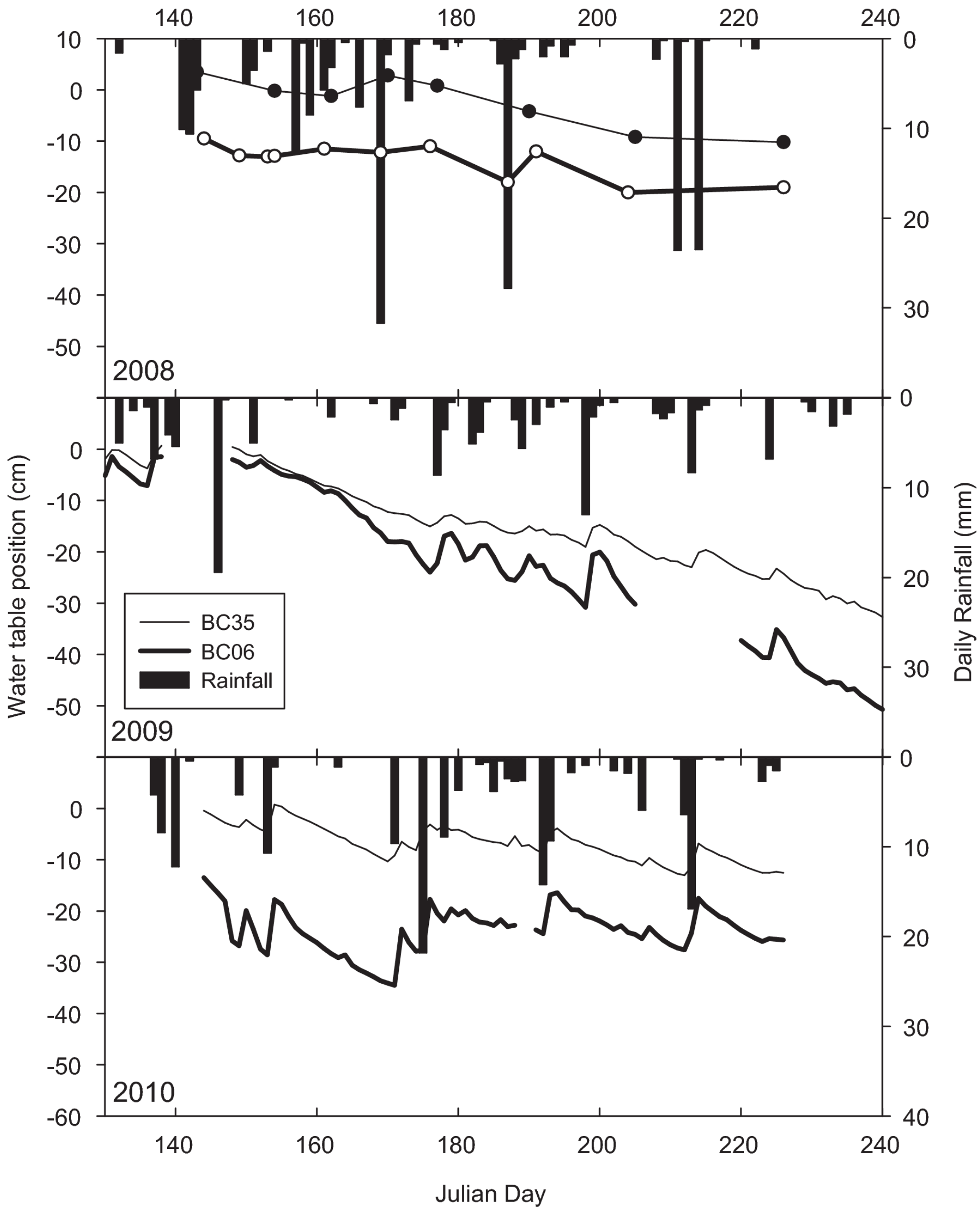

Figure 3.1. Water table and daily precipitation at peatlands recently burned (BC06) and 75 years since burn (BC35). The water table datum for both sites is the surface of an average hollow. Water table data from 2008 represent point measurements with linear interpolation, while 2009-2010 data are daily measurements.

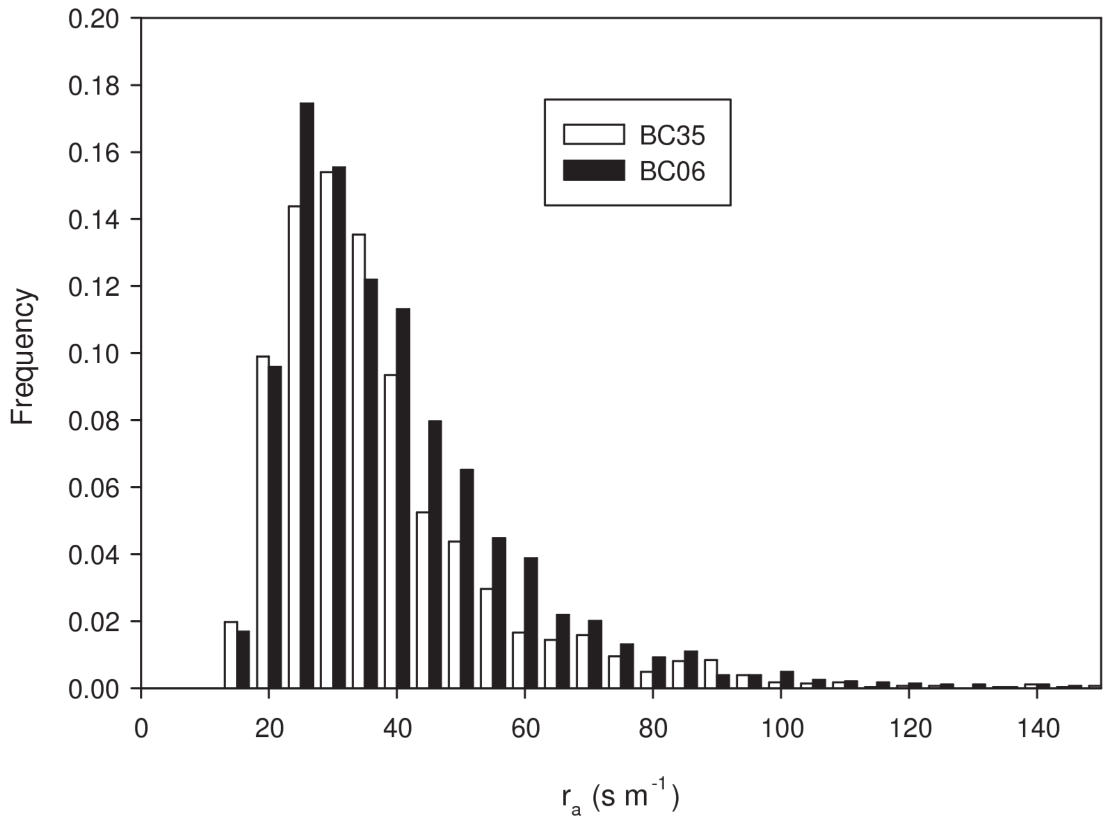

Figure 3.2. Histogram of aerodynamic resistance (r_a) as measured by the wind-profile method for periods, when the wind speed exceeded 1 m s^{-1} 10 m elevation.

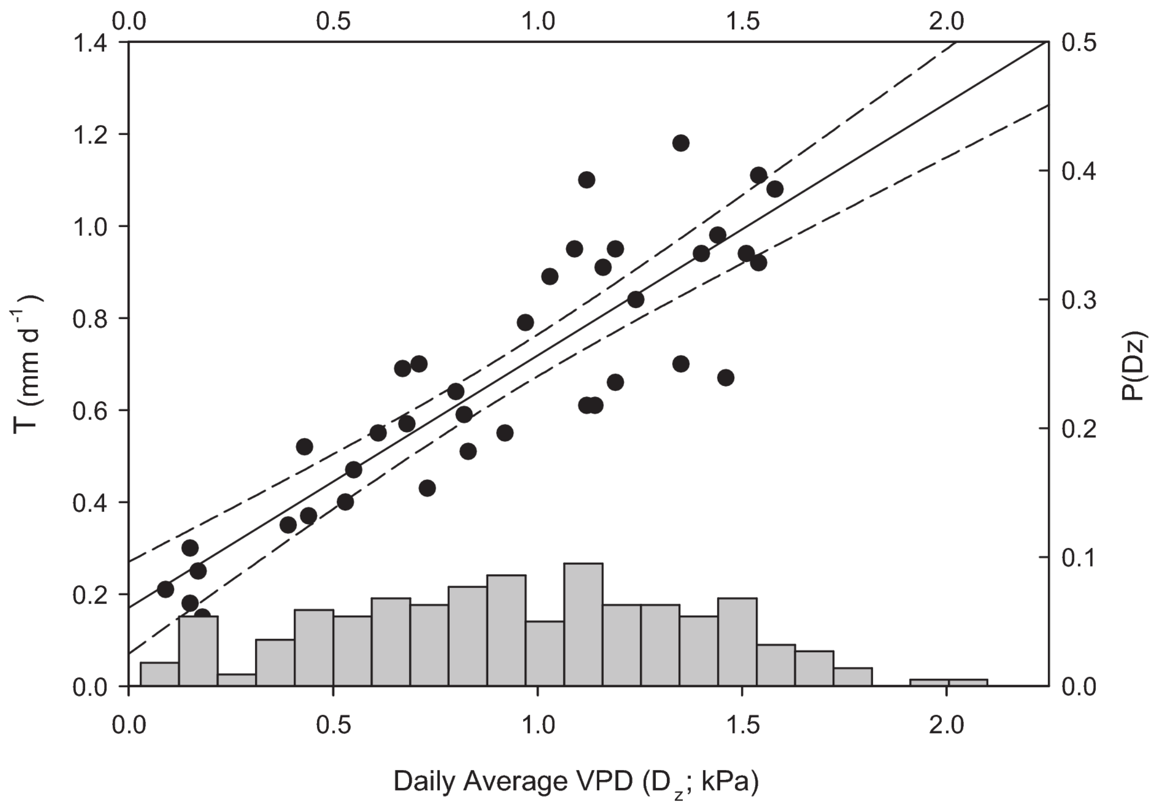

Figure 3.3. Relationship between measured daily transpiration (mm d^{-1}) and daily average vapour pressure deficit (D_z) during 2010. Dashed lines represent 95% CI of the regression line. The frequency distribution of D_z during the period when the sapflow sensors were deployed is shown by the bars along the x-axis.

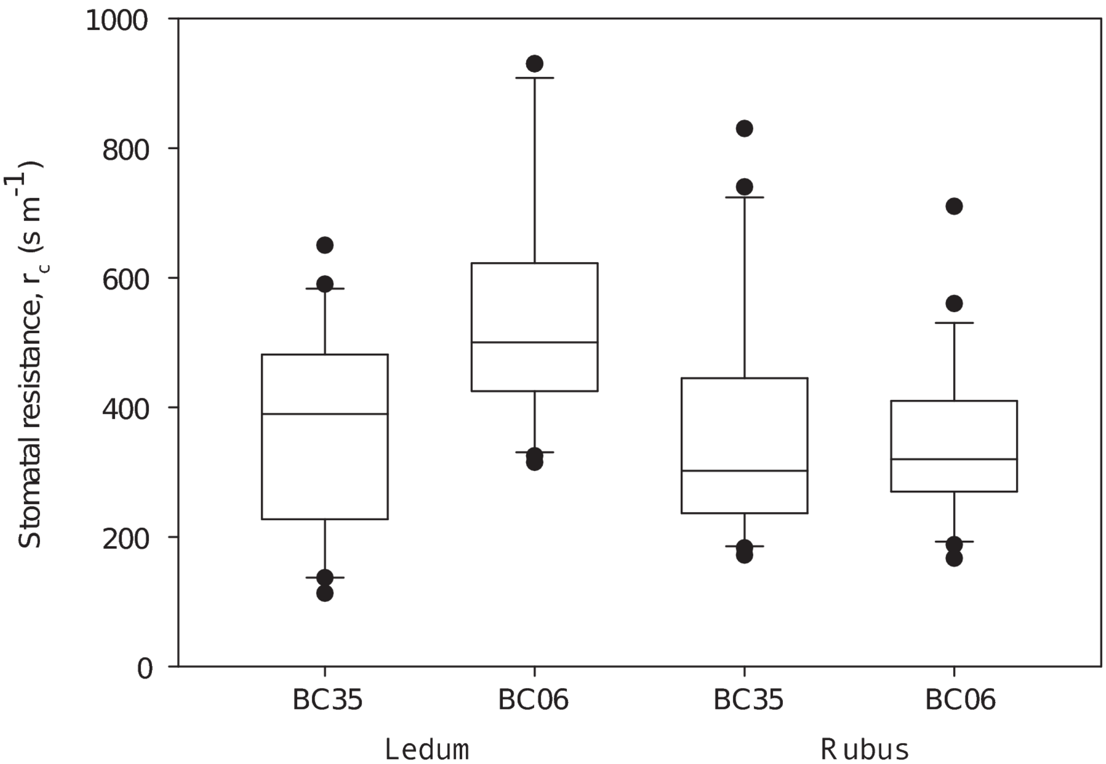

Figure 3.4. Boxplots of stomatal resistance (r_c) in *Ledum groenlandicum* and *Rubus chamaemorus* in both BC06 and BC35.

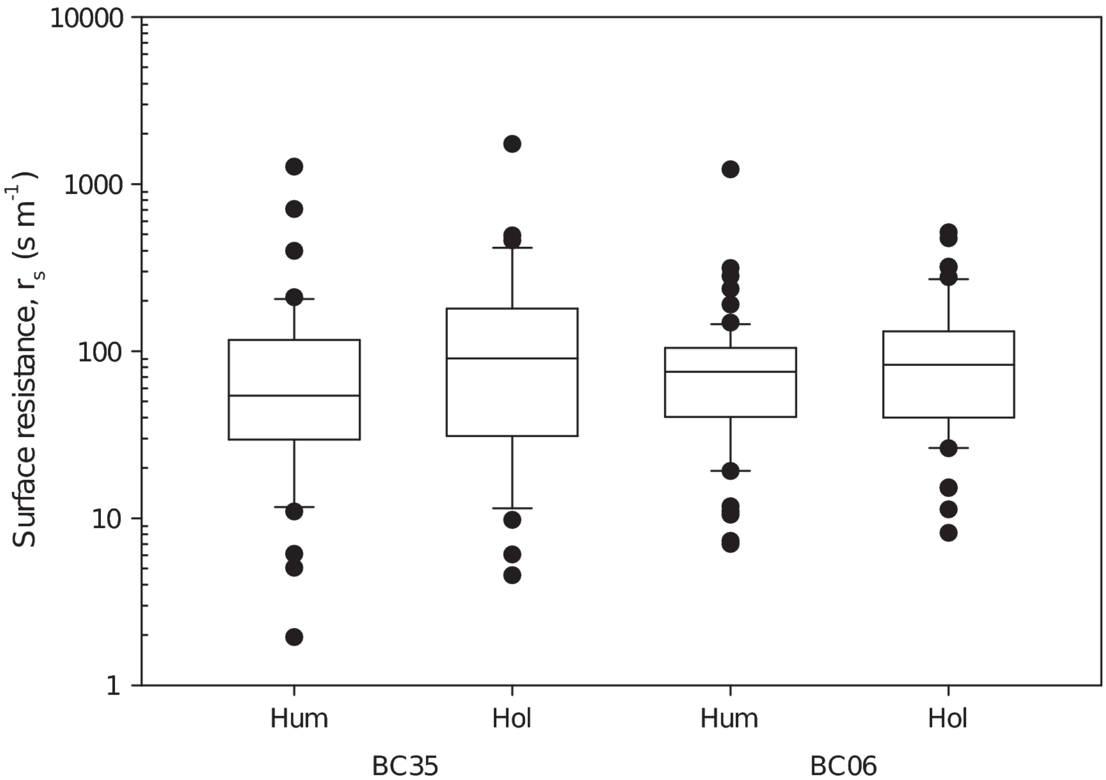

Figure 3.5. Distribution of surface resistance (r_s) measurements as a function of microtopography and site.

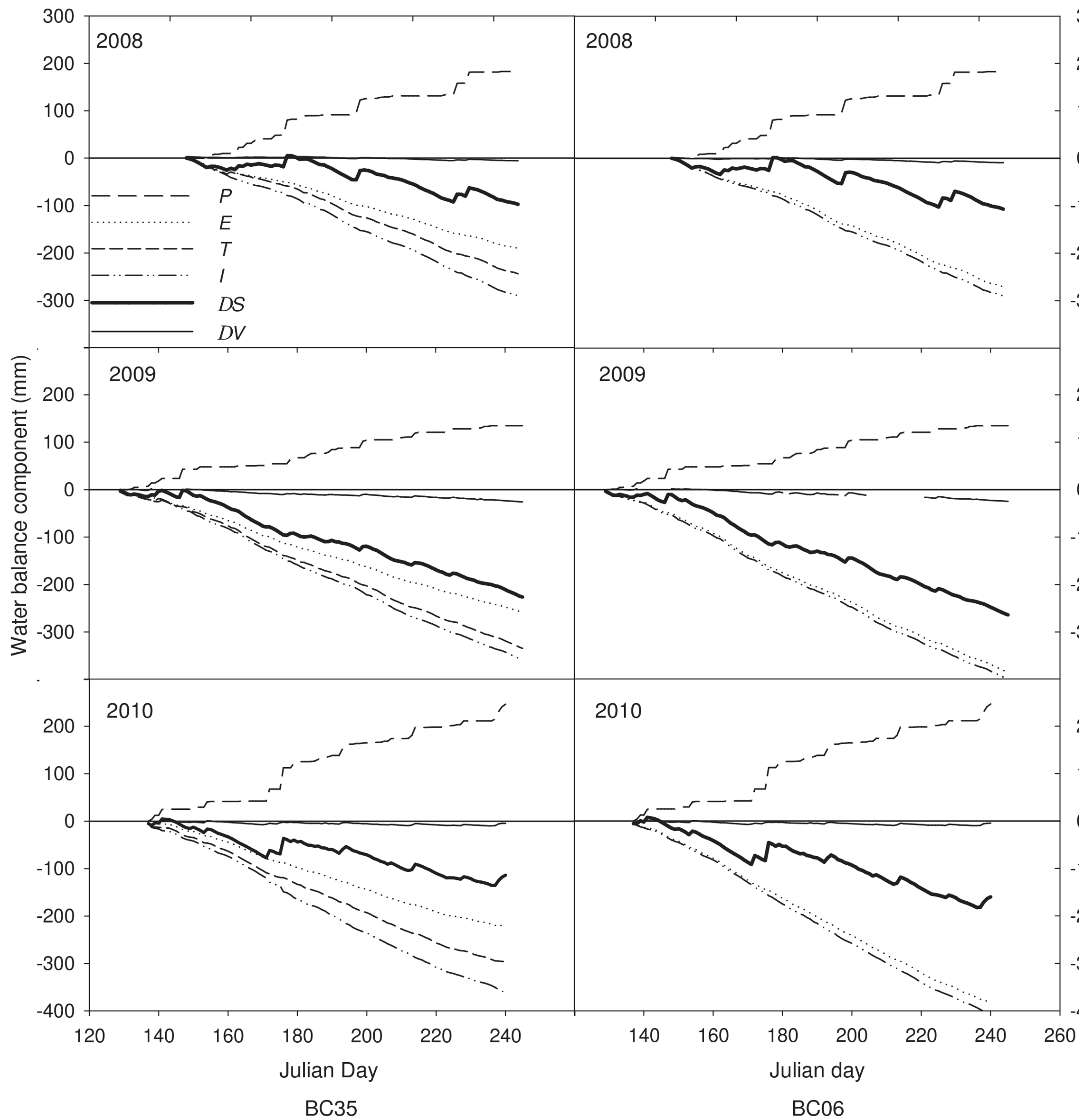

Figure 3.6. Cumulative seasonal water balance of the burned and unburned peatlands from 2008-2010.

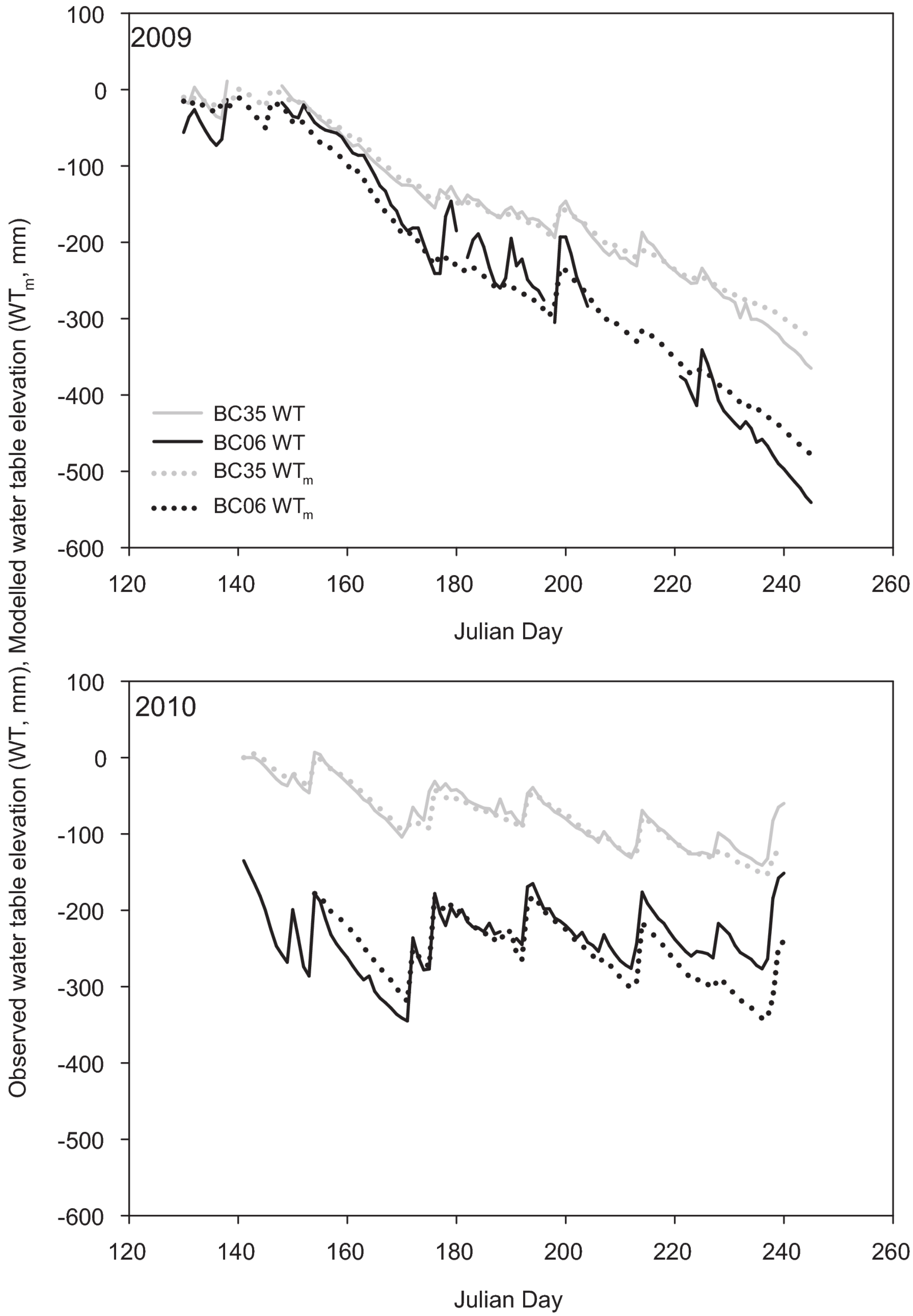

Figure 3.7. Modelled water table (as derived from ΔS) plotted against measured water table at BC35 and BC06 during 2009 (a) and 2010 (b).

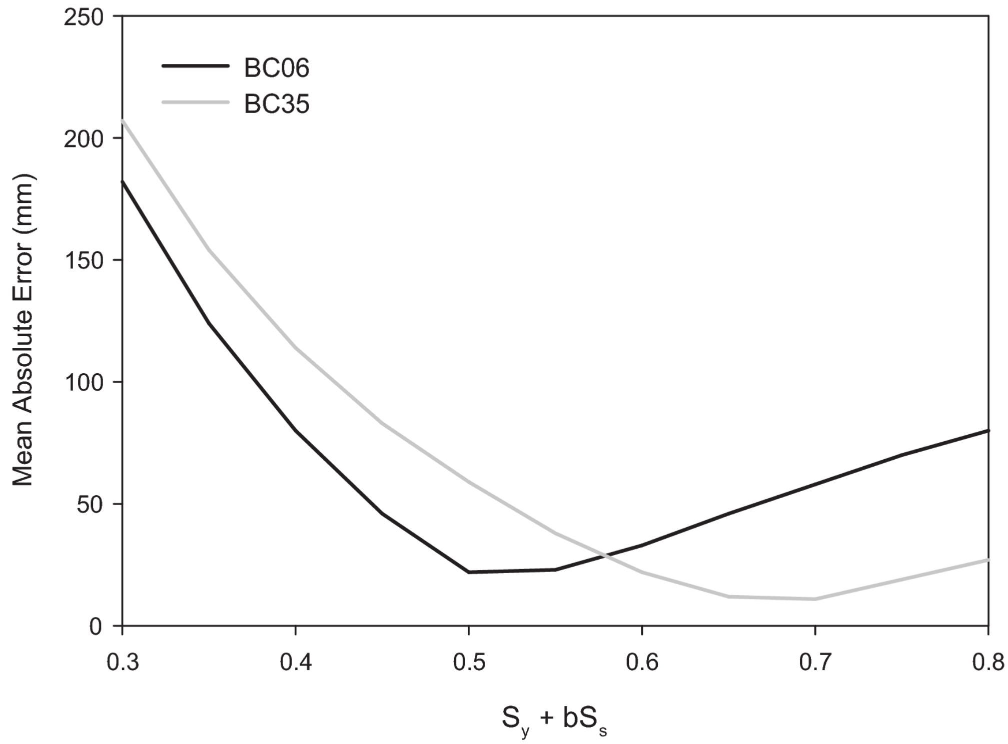

Figure 3.8. Effect of varying specific yield (S_y) on the accuracy of modelled water table in 2009, as measured by mean absolute error, in mm.

CHAPTER 4: EFFECT OF WILDFIRE ON PEAT HYDROPHYSICAL PROPERTIES

4.1. Abstract

We analyzed peat properties and water retention from cores from recently burned and over 75 years since burn forested peatlands in northern Alberta . Bulk density and water retention was greater overall in hollows compared to hummocks, a trend observed at both peatlands. The primary outcome of fire was the exposure of denser peat at the post-fire peat surface, a trend observed more in hollows than hummocks. Humified peat, though it has greater water retention in terms of volumetric water content at a specified pressure of -100 mb, had a lower water retention per unit dry mass. This trend suggests that peat which is more humified may be more vulnerable to smouldering combustion, since there is a greater ratio of fuel (peat) to energy sink (water) given the same hydrological boundary conditions. Fine root content was found to significantly reduce the water retention capacity of peat at a pressure of -100 mb. Fine roots may also retain higher water contents than peat during dry periods, though the same roots can become much drier than the surrounding peat with the subsequent death of trees and shrubs after wildfire. Water retention in peat has implications for post-fire *Sphagnum* regeneration, as more dense peat requires smaller volumes of water loss before a critical growth-inhibiting pore-water pressure of -100 mb is reached. Simulations of water retention after fire showed that hollows are at a higher risk of losing medium-density surface peat, which moderates water table fluctuation and is favourable for *Sphagnum* regeneration.

4.2. Introduction

Peatlands cover approximately 30% of the boreal ecozone across North America, with an estimated carbon stock of 180 Pg (Bridgham *et al.,* 2006). The largest contemporary disturbance to peatlands is wildfire (Turetsky *et al.,* 2002) where the predominantly smouldering combustion of peat can consume in excess of 3 kg C m^{-2} of near-surface peat during a fire event (Benscoter and Wieder, 2003). Given that peat deposits typically store 100-200 kg m^{-2} in total (Beilman *et al.,* 2008), this wildfire disturbance represents a significant short-term carbon flux to the atmosphere, but under normal conditions does not often endanger this carbon pool as a whole. In continental peatlands with a microtopography of hummocks and hollows, field studies have shown depth of combustion is greatest during wildfire in the low-lying hollows, while hummocks remain largely unburned except for the loss of the photosynthesizing capitula (Benscoter and Wieder, 2003). *Sphagnum* mosses comprise upwards of 70% of surface vegetation in peatlands (Benscoter and Vitt, 2008), while the remains of *Sphagnum* are estimated to compose 60% of all peat found in the boreal forest (Zoltai *et al.,* 2000). Recent research suggests *Sphagnum* moss species common to hollows decompose at a much higher rate than hummocks (Turetsky *et al.,* 2008), leading to more dense peat which is associated with enhanced water retention capacity (*e.g.* Boelter, 1968). This increase in ρ_b can also be considered an increase in the fuel available for smouldering combustion during wildfire. The greater consumption in hollows (and therefore generally drier peat) suggests that differences in water retention capacities of differing peat types exert a control on the depth of burn during periods of prolonged drying, where hollow peat may be less able to retain water compared to hummock peat. Standard environmental measurements such as water table (*WT*) depth or volumetric water content (θ) are only indirect indicators of *Sphagnum* moss water retention and drying, as it is the pore-water pressure (Ψ) which dictates drying rates *(Thompson and Waddington*, 2008*)*. The relationship between Ψ and θ, known as the water retention curve, is the standard metric of water retention ability in peat soils (Boelter, 1968; Hayward and Clymo, 1982; Weiss *et al.,* 1998).

To date, no study has systematically examined the controls on water retention in peat soils in the context of microtopography, nor the direct impact of wildfire on peat water retention properties. Weiss *et al.,* (1998) examined the physical controls on peat water retention in *Carex*-dominated peatlands in Finland, and while they only examined peat from below the rooting zone, bulk density (ρ_b) and the proportion of *Sphagnum* and sedge in the peat were found to be the best predictors of water retention. However, this study neither accounted for the presence of microtopography (and the associated *Sphagnum* species contrasts) nor charcoal and other physical impacts of smouldering combustion on a peat profile. Inclusion of microtopography into water retention models can simplify models by narrowing model inputs to known vegetation and peat assemblages, given the large contrasts in ecosystem structure and function, including depth of burn, between hummocks and hollows (Benscoter and Weider, 2003). Moreover, in smouldering combustion the ratio of energy sources (*i.e.* ρ_b) to energy sinks (θ), information provided by the water retention curve, largely determines the depth of burn (Benscoter *et al.,* 2011). Therefore, an increased understanding of the relationships between ρ_b and water retention amongst hummocks and hollows will enhance both the accuracy of combustion modelling and carbon dynamics in peatlands (*e.g*. Turetsky *et al.,* 2002).

Recolonization of *Sphagnum* mosses following wildfire begins within one year and is largely complete within 25 years (Benscoter and Vitt, 2008). While the factors controlling the spatial and temporal variation in recolonization remain unknown, studies in human-disturbed cutover peatlands have shown that *Sphagnum* mosses will not recolonize on a peat surface where the soil-water pressure exceeds -100 mb (Price and Whitehead, 2004). Peat structure and physical properties therefore also play an important role in determining the frequency with which the -100 mb threshold occurs. For example, cutover peat with a higher water retention capacity requires only a small amount of water loss via evaporation before the threshold pressure of -100 mb is attained (Schlotzhaur and Price, 1998). Thus, while high water retention in peat is important for limiting peat decomposition, it is not a desirable quality for the colonization of *Sphagnum* mosses. Instead, peat of lower water retention capacity is preferable, as the hydrological “buffer” of water content prior to reaching the pressure threshold is higher (Figure 4.1). Cut-over peatlands show a dramatic increase in water retention and surface ρ_b due to the removal of less dense surface peat and subsequent compression (Kennedy and Price, 2005). The loss of low-density peat in these systems dramatically decreases the specific yield (S_y) of near-surface peats, causing a more rapid water table decline during rain-free periods (Price and Schlotzhaur, 1999). In peatlands subject to wildfire, the removal of upwards of 20 cm of peat via smouldering combustion is possible (Zoltai *et al.,* 1998), so somewhat similar changes in peat properties to cutover peat may be expected. The hydrological properties of the peat surface after fire has been postulated as a partial control over *Sphagnum* recolonization in the post-fire environment, where hollows regenerate a *Sphagnum* moss cover at a greater rate compared to hummocks (*Benscoter,* 2006). Moreover, in a greenhouse study Campeau and Rochefort (1996) found that regeneration rates of hollow species of *Sphagnum* are enhanced by shallow water table depths, while hummock species such as *S. fuscum* grow at a slower rate than hollow species regardless of hydrological conditions. However, no study has contrasted the post-fire peat properties of hummocks and hollows to determine if physical peat properties contribute to the observed ecological trends.

Here we examine the water retention properties of peat from both a recently burned peatland and one with upwards of 75 years since fire. The objectives of this study are to (i) establish the controls on water retention in the upper 45 cm of peat in both hummocks and hollows, and (ii) determine if wildfire alters the water retention characteristics of the peat surface and how this relates to *Sphagnum* recolonization after fire.

4.3. Site Description and Methods

Triplicate peat cores from hummocks and hollows were collected at a burned and an unburned site from an ombrotrophic peatland 70 km north of Slave Lake, Alberta, Canada in August of 2010. The unburned site was last subjected to wildfire approximately 75 years prior, and the burned site four years prior in September 2006. Peat cores up to 45 cm deep were collected in 10 cm diameter PVC pipes. Hummocks at both sites were primarily *Sphagnum fuscum*. Hollows were a mix of *S. angustifolium*, lichens, and bare peat, while at the burned site they were composed of a bare peat and char mixture at the surface with a sparse cover of *Polytrichum strictum*. Following sampling, cores were frozen for storage and later sectioned into 5 cm thick samples while frozen using a band saw.

Samples were saturated in de-aired water for 24 hours prior to placement on a Soil Moisture Equipment Corp (Goleta, CA) ceramic plate with an air-entry pressure of 1 bar. Samples were placed in PVC pipe sections 10 cm in diameter, and the bottom of the sample was secured with cheese cloth, which allows for an uninterrupted connection between the peat and the ceramic plate (Klute, 1986). The plate was kept in a Perspex box with open water containers adjacent to the ceramic plate, in order to maintain a near-saturated humidity within in the box. While changing pressure steps, the ceramic plate and the interior surfaces of the box were misted with deionized water to further maintain saturation. A central vacuum reservoir was maintained at -300 mb, and a total of four ceramic plates were simultaneously connected to the reservoir via a common pressure manifold with a resolution of 5 mb. Water retention was observed at -10, -20, -30, -40, -50, -75, -100, -150, and -200 mb. Samples were kept at a pressure step for at least 24 hours or until an equilibrium mass was reached, whichever was longer. Water retention was modelled using a version of the Van Genuchten equation (Weiss *et al.,* 1998) that uses a residual water content equal to zero:

$$\theta = \theta_S \left[1 + (\alpha \Psi)^n \right]^{-1 + 1/n} \tag{4.1}$$

where θ is the volumetric water content (m^3 m^{-3}), θ_s is the saturated water content (equal to porosity), Ψ is the pore-water pressure, and α and n are fitted parameters.

Von Post's proportions of fibric material (*F*), fine roots less than 2 mm (*R*), wood and coarse roots (*V*) content and decomposition scale (*H*) (Stanek and Silc, 1977) were used to classify the peat alongside a macroscopic classification of peat botanical origin (hereafter referred to as "peat type"). The variables *F*, *R*, and *V* each range from 0-3, representing quartiles of the volumetric abundance of the material (*i.e.* 25-50% of the sample volume is composed of fine roots where $R = 1$). Porosity of the peat samples was calculated using a particle density for Albertan peat of 1.50 g cm^{-3} (Redding and Devito, 2006). θ_s was calculated to be equal to porosity. Model parameters were fitted using the nls function in the R statistical programming language (R Development Core Team, 2011).

Volumetric water content at -100 mb (θ_{-100}) pressure was measured and used as a metric of moisture retention due to its ecophysiological significance in *Sphagnum* regeneration (Price and Whitehead, 2004) and position near the asymptote of the water retention curves. Furthermore, gravimetric water content ($g_{water}\ g^{-1}_{solid}$) was calculated for samples at a pressure of -100 mb (GWC_{-100}) using θ_{-100} and ρ_b data. Since the samples were the same volume and ρ_b has a known first-order control over water retention (Boelter, 1968), GWC_{-100} is effectively the mass of water retained per unit ρ_b at a constant pressure, or a water retention efficiency. While the strict definition of specific yield (S_y) refers to the water yield per unit head decline as a constant value regardless of Ψ (Hillel, 1996), here we use S_y in the manner common to peatland hydrology, where S_y is the volume of water lost between saturation the first measurement of θ in an unsaturated state, equivalent to the freely drainable porosity (Rydén *et al.,* 1980). In this case, we measured the difference between saturation and the first pressure step (-10 mb) in the water retention laboratory analysis. *Boelter* (1965) and others use free drainage, while Boelter (1968) used -0.1 bars (-100 mb) for analysis of a water yield coefficient similar to specific yield, and potentially a concept similar to field capacity (Hillel, 1996). However, values of free drainage and -0.1 bars were similar between Boelter (1965) and Boelter (1968). In this study, we use not free drainage but a -10 mb pressure.

θ profiles in the peat cores under different water table draw-down scenarios were modelled using the calculated water retention curves for each 5 cm sample and assuming hydrostatic equilibrium, where the pore-water pressure (Ψ) is equal to the depth to the water table (Sumner, 2007). While non-equilibrium conditions are possible in peatlands, they are largely transient in nature (Price, 1997) and require a knowledge of both atmospheric conditions and unsaturated hydraulic conductivity (Price and Whittington, 2010). Change in water storage in the top 50 cm was subsequently calculated as a function of water table depth. Loss of peat by wildfire was simulated by removing the upper peat horizons and adding further peat horizons identical to the deepest part of the sample in order to achieve a depth of 50 cm. It was assumed that no changes occurred to peat water retention properties at or below the burn horizon due to wildfire.

Both θ_{-100} and ρ_b were not strictly normally distributed using a Shapiro-Wilk test of normality largely due to a right skew, but nonetheless approximated the normal distribution sufficiently for Model I ANOVAs to be performed. Von Post's humification scale, *H*, was used as an input to into an ordinary least squares regression model of θ_{-100}, as it approximates interval level data. Fibre (*F*), root (*R*), and wood (*V*) contents were also used in the OLS regression as the scale of 0-3 is also interval level data because each of the scales approximates 0-25, 25-50, 50-75, and 75-100% content of the respective material type.

4.4. Results

4.4.1. Bulk Density

A two-way ANOVA of site (burned/unburned) and microtopography (hummocks/hollows) showed a significant impact of site on ρ_b ($F_{1,94} = 17.5$, $P < 0.001$), with mean ρ_b at the burned site (0.085 g cm^{-3}) greater than that at the unburned (0.056 g cm^{-3}). Moreover, microtopography significantly influenced ρ_b ($F_{1,94} = 51.0$, $P < 0.001$), with ρ_b in hollows (0.095 g cm^{-3}) being greater than in hummocks (0.045 g cm^{-3}). There was a weak interaction between microtopography and site ($F_{1,94} = 3.2$, $P = 0.075$), with higher ρ_b in burned hollows. ρ_b showed only a weak linear dependence on depth from the surface in both hummocks and hollows ($R^2 = 0.31$; $F_{1,96} = 43.7$; $P < 0.001$). Bulk density of core sections increased with depth down the core in 52% of samples in hummocks and 51% of adjacent samples in hollows. However, there was on average a 5% increase in ρ_b between adjacent sections, resulting in the expected gradual increase in ρ_b with depth. Degree of humification (*H*) influenced bulk density ($R^2 = 0.47$; $F_{1,96} = 89.1$; $P < 0.001$), though depth and *H* were themselves correlated ($r = 0.55$).

A one-way ANOVA revealed a significant effect of peat type ($F_{4,93} = 22.7$, $P < 0.001$) on ρ_b. A Tukey's HSD post-hoc test revealed that ρ_b in humified peat was greater than in both *S. fuscum* and root peat. There was an insufficient number of char and *S. angustifolium* peat samples to make any statistical conclusions,, though *S. angustifolium* and surface charred peat tended to be lower than average in ρ_b.

4.4.2. Water Content

Specific yield (Figure 4.2a) and θ_{-100} (Figure 4.2b) were largely a function of bulk density, which as mentioned above, itself was weakly correlated with depth. The relationship between ρ_b and both S_y and θ_{-100} was highly heteroscedastic, with increasing variance as ρ_b increases. The line of best fit from Boelter (1968) shows good agreement with this dataset, though the data presented here show θ values approximately 10% higher for a given value of ρ_b. Influences on θ_{-100} were similar to ρ_b, with site, H, and microtopography playing significant roles in determining θ_{-100} (Figure 4.3). The burned site showed higher overall θ_{-100} ($t_{1,96} = 2.1$, $P = 0.037$). A two-way ANOVA of site and microtopography revealed that unlike ρ_b, where site showed the strongest control, microtopography had the greatest influence in determining θ_{-100} ($F_{1,94} = 54.5$, $P < 0.001$). A one-way ANOVA of θ_{-100} by peat type revealed that peat type strongly influenced θ_{-100} ($F_{4,75} = 109$, $P < 0.001$), though Tukey's HSD test showed that only rooty peat showed any significant difference in θ_{-100} (Figure 4.4). A multivariate ordinary least squares (OLS) regression of θ_{-100} showed that a combination of ρ_b, H, and R were significant predictors of θ_{-100} ($\theta_{-100} = 2.34\rho_b + 0.025H - 0.094R + 0.194$; $R^2 = 0.80$; $F_{6,91} = 63.4$; $P < 0.001$), with ρ_b and R being the strongest predictors. While sample depth was included in the initial regression, it was excluded due to its limited predictive power and correlation with ρ_b. Overall, 80% of the total variance in θ_{-100} was explained by these three variables alone. An examination of GWC_{-100} across all the samples showed that humification appears to decrease GWC_{-100} (Figure 4.5).

4.4.3. Water Retention Model Parameters

The parameter α of the Van Genuchten model describes the convex ($\log_{10}(\alpha) < 0$) or concave ($\log_{10}(\alpha) > 0$) nature of the water retention curve, and was the more varied of the two parameters. Values of α before log transformation varied from 3647 to 0.01. A two-way ANOVA revealed that $\log_{10}(\alpha)$ was influenced by topography ($F_{1,87} = 8.2$; $P = 0.005$), where hummocks exceed hollows, but neither site nor an interaction between site and microtropography was a significant factor. Overall, ρ_b was a poor predictor of $\log(\alpha)$ ($R^2 = 0.30$; $F_{1,89} = 40.4$; $P < 0.001$; Figure 4.5a). For the entire dataset, a robust linear relationship was observed between $\log(\alpha)$ and the specific yield of samples (Figure 4.6).

The parameter n in the Van Genuchten model is an exponent that in part describes the rate at which the water retention curve reaches a horizontal asymptote at more negative pressures. Values of n varied considerably less than α, ranging from 1.0 through 1.9. Mean n across all samples was 1.3, with generally lower and less varied values of n across humified and other more dense samples than in less dense samples (*S. fuscum* in particular), which showed much greater variance. There was no significant correlation between n and $\log_{10}(\alpha)$ ($r = 0.19$; $t_{89} = 1.87$; $P = 0.06$).

4.4.4. Impacts of Wildfire on Water Retention in Surface Peat

The hollow samples featured a variety of peat types in the uppermost sample (0-5 cm). However, the hummock samples offered the opportunity to contrast water retention between living *S. fuscum* moss and undecomposed *S. fuscum* peat remnants after the wildfire burned off just the living moss capitula, (resulting in a depth of burn likely less than 2 cm). At first examination, the water retention curve of the burned hummocks showed one sample with a greater retention capacity than the unburned samples (Figure 4.7a). However, when *GWC* is shown as opposed to θ (Figure 4.7b, d), the higher ρ_b sample from the burned site has a similar retention efficiency, and the remaining two burned samples have water retention efficiencies lower than unburned samples. In hollows, the largest contrast between samples existed at the unburned site, where θ_{-100} varied from 0.2 to 0.75 (Figure 4.7c). In the case of the unburned hollow sample where θ_{-100} = 0.2, the Von Post score of root abundance (*R*) was 3, meaning the sample was composed of 75-100% fine roots by volume. However, when the water retention curves were normalized for ρ_b (Figure 4.7d), this root-abundant sample appears to have a similar water retention efficiency to samples with less fine roots. Overall, the majority of both the surface hummock samples and hollow samples at depth have a *GWC* between 500% and 700% at -200 mb suggesting this metric has promise in identifying trends in water retention that are not due to differences in ρ_b.

In order to simulate the impacts of the removal of the surface layer on changing ecohydrological thresholds for *Sphagnum* recolonization, changes in θ_{-100} were plotted against depth for the six cores collected from the unburned site (Figure 4.9). Following a wildfire resulting in a depth of burn of 20 cm, hollows showed a larger change in θ_{-100}

than hummocks, increasing by between 0.1 and 0.5, with considerable variation between cores. Hummocks, however, showed both a smaller increase in θ_{-100} than hollows (0.02-0.19), but also far less variation between cores and as a function of depth of burn. The loss of near-surface peat can also be visualized as the amount of change in storage (via evaporation or runoff) required to reach a given water table depth (Figure 4.10). Prior to fire, the hollow modelled here required 180 mm of net water loss before a water table of 100 cm was reached, corresponding to a pore-water pressure of -100 mb at equilibrium. Loss of the upper 20 cm during wildfire would reduce the required ΔS roughly by half, so that only 95 mm of water loss is required to reach a 100 cm water table. In contrast, the less dense peat in the hummocks required 390 mm of water loss to reach 100 cm water table depths, while the combustion of 20 cm of peat would decrease this amount to 360 mm.

4.5 Discussion

4.5.1 Physical Properties

Bulk density showed a similar topographic contrast at both sites, with hollows being more dense compared to hummocks, particularly at depths greater than 10 cm. The lower density of *S. fuscum* dominated hummocks is consistent with decomposition studies (*e.g.* Johnson and Damman, 1991) that show *S. fuscum* as having slower decomposition rates (and hence lower ρ_b) even in reciprocal transplants, suggesting the physical properties of *S. fuscum* and not the soil environment are the cause. Turetsky *et al*. (2008) demonstrated that the carbohydrate structure of *S. fuscum* is largely responsible for the decrease in decomposition rates compared to hollow species of *Sphagnum*. This microtopographic effect on ρ_b was also shown in the correlation between ρ_b and humification, where the frequency of moderate to highly humified peat ($H > 7$) was less in hummocks.

Though an overall trend towards increasing ρ_b with depth was observed across all cores, the trend was not monotonic. The observation that just under half of all adjacent depth-wise sections showed a small decrease in ρ_b down the core is exemplary of the non-linear nature of peat profiles, where sharp contrasts in peat properties can occur with depth. Benscoter *et al.* (2005) showed that hummocks often revert to hollows and vice versa after wildfire, thus creating a hummock-hollow contrast vertically within a peat profile, and the bulk density contrasts therein. Overall, there was a strong correlation amongst depth, H, and ρ_b in the data presented here, similar to the relationship found for Scandinavian peats by Paivanen (1973). Stanek and Silc (1977) observed a very strong correlation between ρ_b and H ($r = 0.94$) for peat in forested peatlands in northern Ontario, Canada, though no microtopography was noted. The strong microtopographic contrast in ρ_b suggests that datasets of peat properties without species or microtopographic information, such as Zoltai *et al.,* (2001) and Shaw *et al.* (2005) are more limited in their utility and predictive capacity in peatlands with a strong microtopographic gradient.

4.5.2. Water Retention Properties

Volumetric water content in peat has been previously shown to be largely controlled by ρ_b, a trend first formally identified by Boelter *(*1968*)*, who noted a shift of θ_{-100} from 0.15 to 0.75 between undecomposed surface *Sphagnum* and well-decomposed peat. While Paivanen (1973) found a curvilinear relationship between ρ_b and θ_{-100}, our data show instead a linear relationship with strong heteroscedacity, with variance and residuals increasing with increasing ρ_b (Figure 4.2). A θ_{-100} of approximately 0.8 is achieved in samples with a ρ_b between 0.07 and 0.20 g cm^{-3}, showing that peat properties outside of ρ_b can exert a strong control on θ_{-100}. This variance in θ_{-100} is almost entirely right-skewed, however. No outliers in Figure 4.2b lie to the left of the main cluster of data, such that all deviations from the relationship lie in a lower θ_{-100} for the same ρ_b (or higher ρ_b for the same θ_{-100}). In effect, there is a line of optimal water retention for a given ρ_b, dominated by *Sphagnum* peats, on the left-hand side of the cluster of data. Roots, char, and humified peats serve only to decrease water retention, and are shown by plotting further to the right of this line. Gravimetric water content at a given pressure shown in Figure 4.8 effectively normalizes for variations in ρ_b. Roots, wood, and other non-*Sphagnum* materials have a lower GWC_{-100} (Figure 4.8) compared to *Sphagnum* peat. While the surface peat, though it has a high S_y is typified by a very small difference between θ_{10} (the pressure at which S_y was measured) and θ_{-100}. Once this low-density peat has been unsaturated, only a small volume of net water loss is required in order to reach a pore-water pressure of -100 mb.

Peat with high root content appears to have lower water retention when compared to *Sphagnum* peat of similar ρ_b. While the data shown here is for the laboratory environment where the roots have been severed from the overlying vegetation, consideration of roots in water retention curves is complicated by their active transport of water while living, a process that would otherwise enhance root water contents. For that reason, many authors such as Weiss (1998) have avoided roots, and instead sampled exclusively below the rooting zone. However, given the importance of wildfire disturbance on near-surface peat, root water content should be considered. While there are no existing studies of root water content in peatlands or in ericaceous shrubs, Bois *et al.,* (2006) measured a fine root *GWC* of ~400% in *Picea glauca* potted seedlings. In comparison, a sample in this study almost entirely composed of fine roots (R = 3) from a depth of 15-20 cm in an unburned hollow had a GWC_{-100} of 72%, indicating that water retention analysis likely significantly under-reports water contents in peat with abundant fine roots. However, water retention analysis is still relevant and presumably more accurate in peat after wildfire disturbance, where shrub die-off after fire renders the roots dead and solely passive water retention medium. As a result, in the post-fire environment rooty peat horizons hold less water than *Sphagnum* peat (Figure 4.4). High fine root content has been observed in the surface of burned hollows immediately after fire (N. Kettridge, pers comm.), suggesting that root mats help promote high water content in peat during dry periods and potentially suppress smouldering combustion. A root *GWC* of approximately 400%, while lower than the *Sphagnum* of 600-800% (Figure 4.7b), is significantly higher than that of feathermosses, which routinely decline to a *GWC* of less than 100% during rain-free intervals of one-week or more (Wilmore, 2001). Thus, roots play a moderating role in the water retention of a peatland, simultaneously lowering water content compared to *Sphagnum* peat, but increasing water retention compared to the drier, shade-adapted feathermosses typical of late successional forested peatlands

(Benscoter and Vitt, 2008) that typically burn at 100-120 year intervals (Turetsky *et al.,* 2004).

4.5.3. Impact of Fire on Water Retention

Previous investigations of the impacts of wildfire on soil properties have focused heavily on the mineral soils of forests where the overlying layer of duff and litter as well as the organic portion of the soil is consumed (*e.g.* DeBano, 2000). However, in the case of peatlands, the combustion of a deep, purely organic soil presents a variety of distinctions. Foremost is the loss of the pre-fire soil elevation datum. While the depth of burn can be estimated using the adventitious root method (Kasischke *et al.,* 2008), the exact depth of burn in a particular location cannot be known under field conditions without a having a burn pin or other instrumentation in place prior to the fire. Given this limitation, we have chosen to use the peat surface upon sampling as the zero depth datum, while in fact this elevation would have been at depth prior to the fire. Since peat generally follows a trend of increasing ρ_b with depth (Zoltai *et al.,* 2000), the effect of fire is therefore to increase bulk density on average purely due to a shift in the surface datum. This effect is enhanced by a tendency for smouldering fronts to halt at points of abrupt vertical increase in ρ_b due to energetic limitations, as shown in laboratory studies by Benscoter *et al.* (2011).

The hummock surface composed of *S. fuscum* provided a scenario most similar to mineral soils in that there was limited vertical depth of burn, in this case modification of remaining soils by fire and heating is of interest. The capitula remained partially intact after fire, and only a small depth of burn on the order of 1-2 cm was likely. Hummocks of *S. fuscum* showed less variability in density and humification with depth and lacked visible charcoal or ash. By examining the GWC_{-100} to adjust the samples for differences in ρ_b, two of three surface samples of *S. fuscum* from the burned site showed a decrease in GWC_{-100} compared to unburned samples at the same tension (Figure 4.7b), suggesting an impact of fire on water retention, outside of changes in bulk density. This effect may have been brought about by heating and subsequent alteration of organics in the moss itself, an effect seen widely in mineral soils with significant organic content (e.g. Stoof *et al.,* 2010).

Hollows are more difficult to compare directly, given their greater variability in peat type and density, as well as an unknown depth of burn. Accordingly, hollow samples at the same depth do not show a difference in water retention once differences in bulk density are accounted for (Figure 4.7d). Notable in hollows, however, is the presence and preservation of charcoal, which has been shown to have inferior water retention properties as compared to mineral soils (Stoof *et al.,* 2010). At the surface of the burned hollows, a charcoal layer on average 1-2 cm deep was observed, and this corresponded to a lower water retention in terms of both θ and GWC compared to samples directly underneath with no significant char content.

The role of increasing depth of burn on post-fire water retention is largely a function of the bulk density profile with depth. Given the generally increasing bulk density with depth, greater depth of burn typically leads to greater moisture retention (*e.g.* θ_{-100}). The effect is most pronounced in hollows, where the bulk density gradient is largest. Diminishing gains in water retention present themselves in hollows where the depth of burn exceeds 10-15 cm (Figure 4.9), as the largest contrasts in density occur in the near-surface layers. A more linear relationship is observed in hummocks, where increased depth of burn is accompanied by similarly large increases in moisture retention due to the more uniform peat bulk density profile. In cases of exceptional depth of burn (>30 cm) where a deep water table is present, the only advantage to a high depth of burn may be the physical lowering of the peat surface datum relative to the water, thus lowering the expected post-fire pore-water pressures.

4.5.4. Implications for Post-fire Sphagnum Colonization

Ultimately, the water retention curve and θ of a particular peat surface is only of minor importance for *Sphagnum* recolonization, as it is pore-water pressure and not θ itself that is the potentially limiting factor (Thompson and Waddington, 2008). θ at the surface is only the expression of the water retention curve, although the retention curve does become important where the volume of water required for water table draw-down to 100 cm (Figure 4.10) is of interest. Since the depth to water table determines the pore-water pressure at the surface (at least in equilibrium scenarios), the water retention curve only indirectly plays an important role in influencing *Sphagnum* recolonization. However, since the water table is generally flat under hummocks and hollows (*Price and Whitehead*, 2004), the importance of S_y at a point – and therefore water retention – is limited as water table behaviour is the aggregate of all peat in the local area. It is therefore important to note the abundance of hummocks and hollows on the landscape in addition to the water retention properties of each microform. Although data are limited on this subject, Benscoter *et al*. (2005) observed a decline in hummock area from 80% to

45% after wildfire, while the observed change at BC06 was smaller decline, from 72% to 62% (Chapter 3).

Despite a smaller difference between θ_{-100} and θ_s in the generally denser hollow peat (Figure 4.2), hollows have higher re-colonization rates compared to hummocks (Benscoter, 2006). The advantage in recolonization held by hollows may not be entirely due to water retention, but also due to the relative position of the hollow surface closer to the water table. Hummocks are typically 20-40 cm taller than hollows, with a maximum height of over 70 cm (Andrus *et al.,* 1983). Since a pore-water pressure of -100 mb (or 100 cm depth to *WT*) is the limiting threshold for *Sphagnum* regeneration (Price and Whitehead, 2004), hummocks are 40-70% closer to the critical pressure by virtue of their elevation alone. The presence of hummocks on the landscape reduces the depth of burn, and therefore minimizes changes to the ΔS-*WT* relationship after fire (Figure 4.10), resulting in less *WT* fluctuation after fire. Hummocks are at a significant disadvantage in terms of regeneration after fire, despite their low depth of burn during wildfire itself. Therefore, hummocks are beneficial to the landscape as a whole in that they maintain a high S_y and minimize *WT* fluctuations, but themselves are at a disadvantage in terms of *Sphagnum* recolonization. A decrease in the abundance of hummocks is a common element in many types of disturbance (Turetsky *et al*., 2011) and subsequent removal of the low density peat may be an important tipping point in the transition toward a low S_y, low *WT*, and ultimately high *h* environment.

4.6. Conclusions

Water retention analysis revealed that ρ_b plays the largest role in determining the capacity of peat to retain water. Hummocks appear to be more resistant to smouldering combustion due to their higher surface moisture content, which is largely due to their greater ρ_b at the surface. Hollows showed high ρ_b at depth, but the surface peat was often very loose and poorly cohesive, resulting in lower water retention at the surface compared to hummocks. Both ρ_b and water retention increased after fire in both hummocks and hollows, though this is likely more an artefact of the combustion of less dense near-surface peat rather than any *in situ* changes to the peat itself. For humified peat, this relationship between ρ_b and water retention did not occur, and greater humification caused less water to be retained per unit dry mass. This finding may have importance in the studies of the thermodynamics of smouldering combustion of peat, as humified peat requires less energy to drive off water prior to combustion for the same mass of fuel.

The presence of abundant dead fine roots was found to significantly decrease both ρ_b and water retention in peat. Prior to fire, however, the same fine roots occurring as part of a living plant may have contributed to higher overall moisture in the peat, and to the cessation of smouldering due to high water content. Given the abundance of fine roots in samples from both burned and unburned sites, studies of fine roots in forested peatlands and particularly their water contents and ability to alter the pore-water pressure, warrants further investigation.

The post-fire peat surface shows some evidence of lower water retention compared to pre-fire peat surfaces, but only once differences in ρ_b are accounted for.

Hollows after fire experience a much larger increase in ρ_b at the surface, which then creates an environment more susceptible to hydrological extremes of drying. Hummocks feature a more uniform peat of moderate density that requires a much larger loss of water in order to lower the water table. By promoting a high S_y, the hummocks serve to resist water table draw-down after fire, which then decreases the probability of pore-water pressures exceeding thresholds known to inhibit *Sphagnum* recolonization. This study provides evidence of how hummocks, particularly those of *S. fuscum*, serve to promote peatland resilience. Specifically, *S. fuscum* prevents the dominance of dense and low S_y peat (though *S. fuscum* has high water retention) that ultimately can lead to peatland decline through a positive feedback of low S_y and water table drawdown.

Ultimately, high water retention does not appear to be an asset in the context of continued peatland ecosystem function, particularly with regards to *Sphagnum* recolonization. High water retention capacity limits the difference between peat at θ_s and at θ_{-100}, where *Sphagnum* stress occurs. This high water retention is also correlated with lower porosity and therefore lower specific yield, which subsequently leads to more rapid water table decline. Peat of moderate ρ_b may therefore strike the best balance between resisting combustion and maintaining a high S_y. While high water retention is a beneficial quality with respect to resistance to smouldering combustion, it is indeed the ratio of energy sinks to sources, as expressed by GWC_{-100}, that is a more suitable metric of smouldering resistance. Future models of forested peatland carbon exchange, ecosystem function, and vulnerability to wildfire will be improved by including a process-based representation of peat water retention and its determinants.

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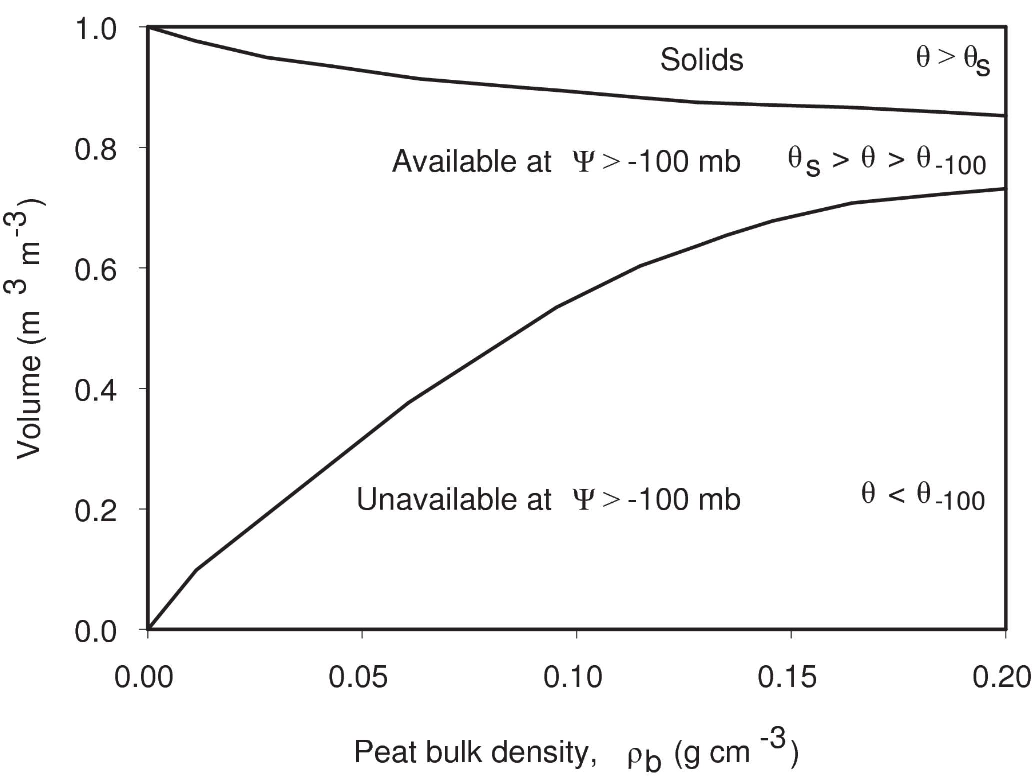

Figure 4.1. Volume of water available at pore-water pressures greater and less than -100 mb as a function of peat bulk density. The uppermost area represents the volume occupied by the organic particle solids themselves. Modified from *Boelter* (1968).

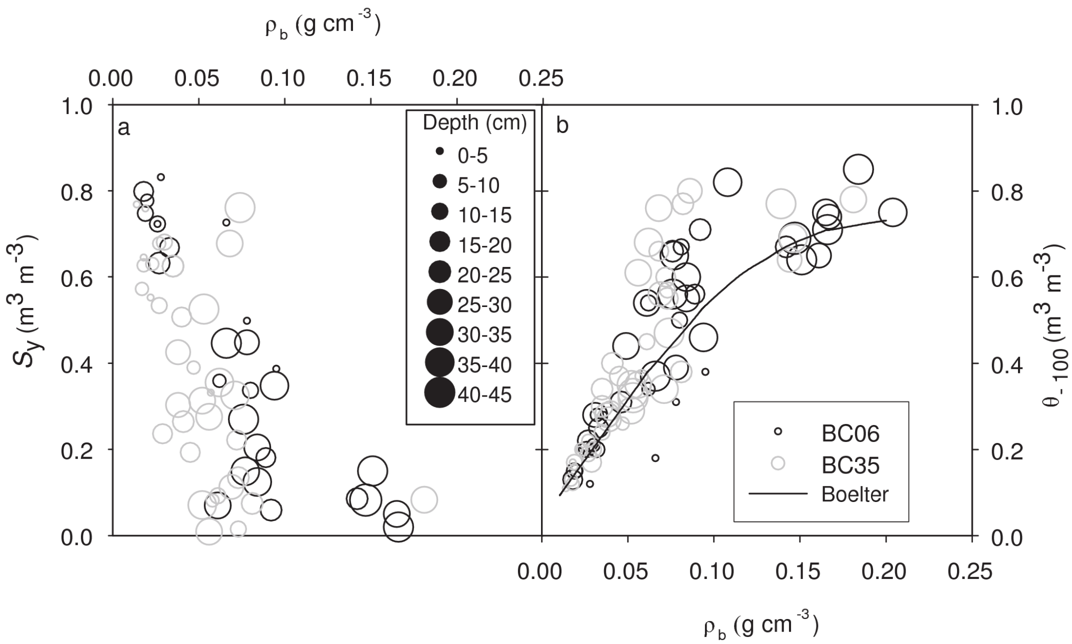

Figure 4.2. Specific yield (S_y; panel a) and volumetric water content (panel b) at -100 mb pressure (θ_{-100}) as a function of bulk density. The relationship between (θ) and ρ_b at -100 mb from *Boelter (*1968*)* is shown as a solid line for reference. Bubbles size is proportional to sample depth, with the same scale for both burned and unburned sites.

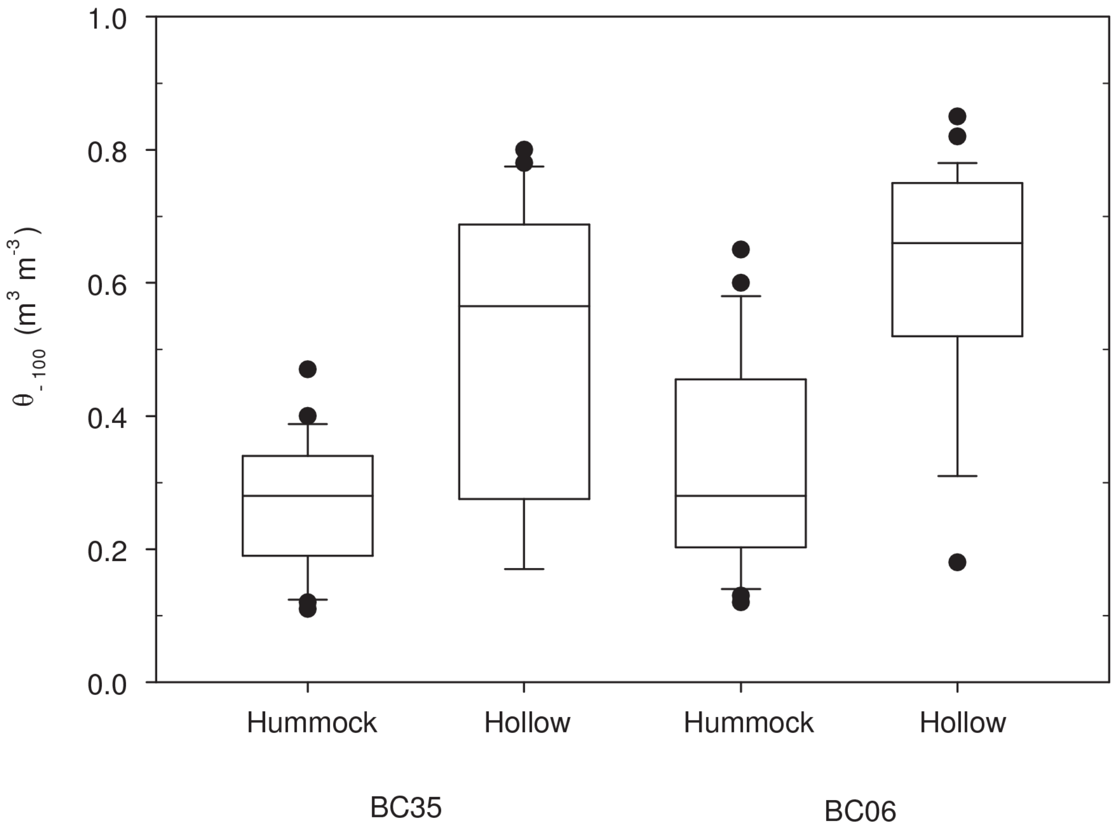

Figure 4.3. Distribution of θ_{-100} amongst hummocks and hollows at the burned and unburned sites.

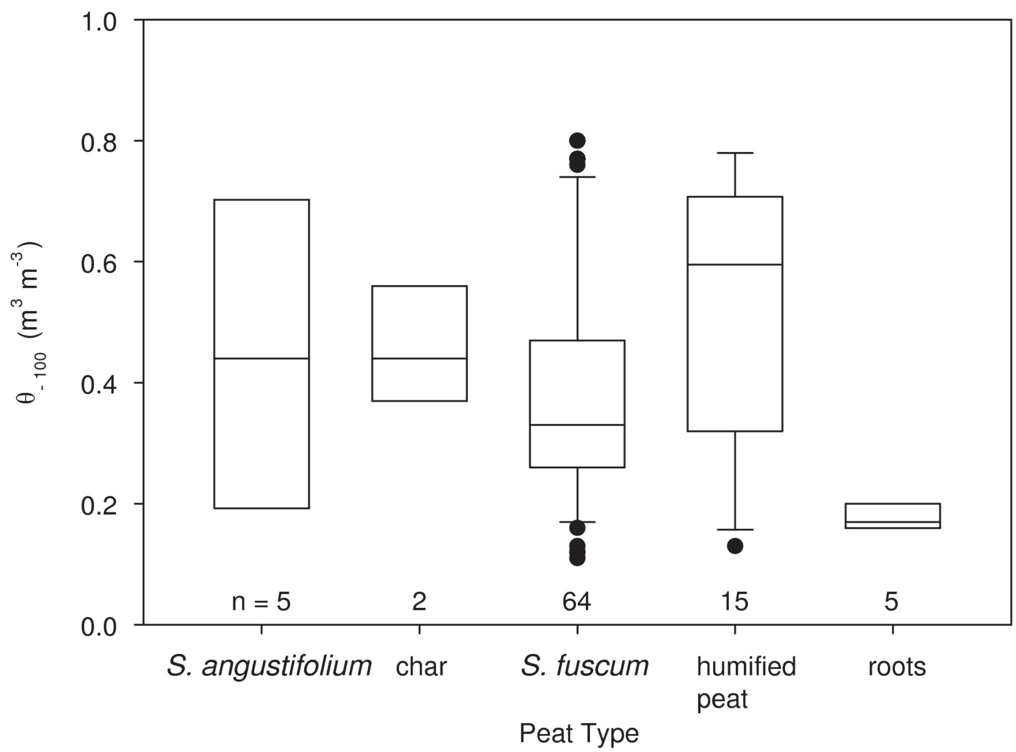

Figure 4.4. Distribution of θ_{-100} amongst differing peat types.

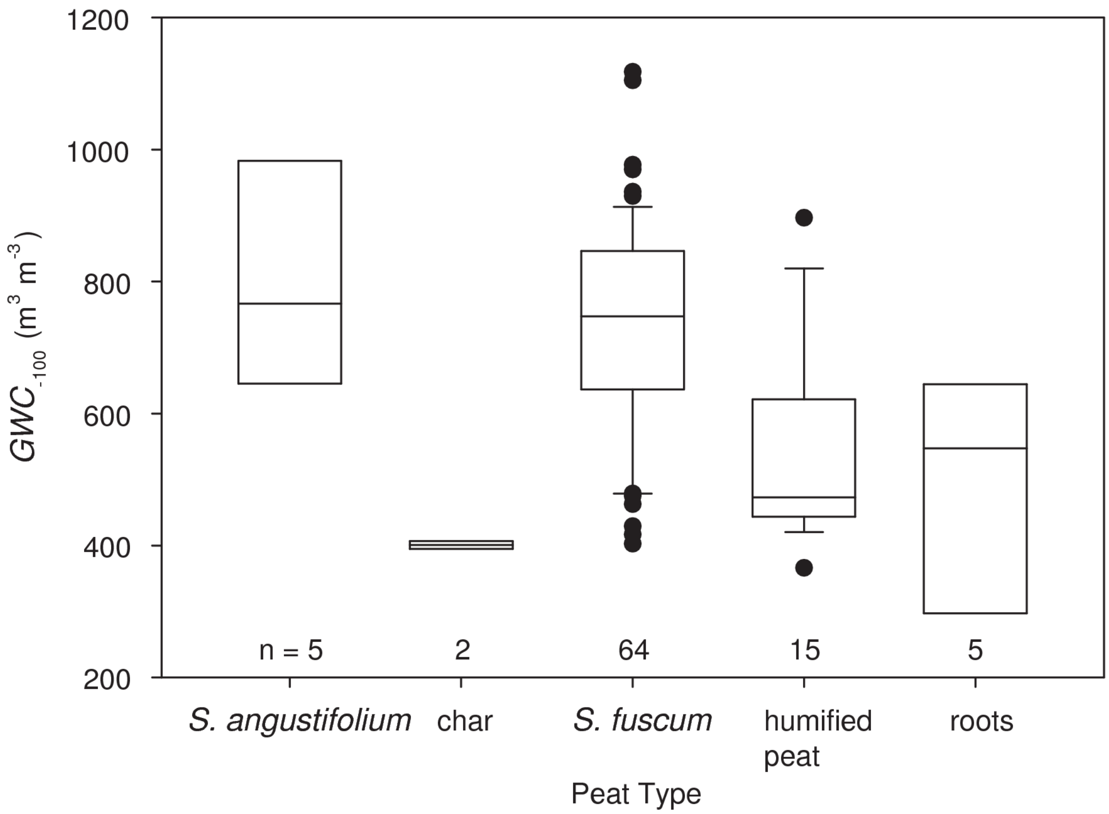

Figure 4.5. Effect of peat type on the efficiency of water retention in peat, expressed as a gravimetric water content at -100 mb pressure (GWC_{-100}).

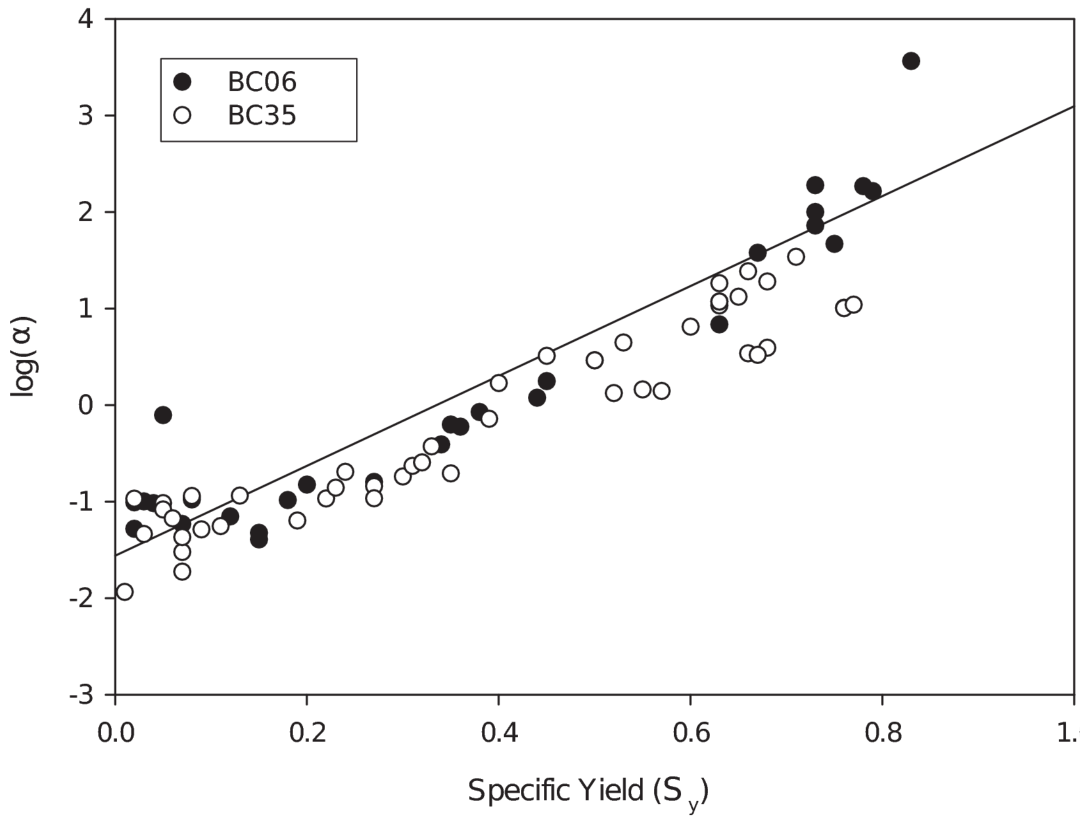

Figure 4.6. Relationship between the Van Genuchten model parameter α and specific yield in peat. The regression line follows the equation: $\log(\alpha)=4.1S_y-1.5$; $R^2 = 0.86$; $F_{1,71} = 441.3$; $P < 0.001$.

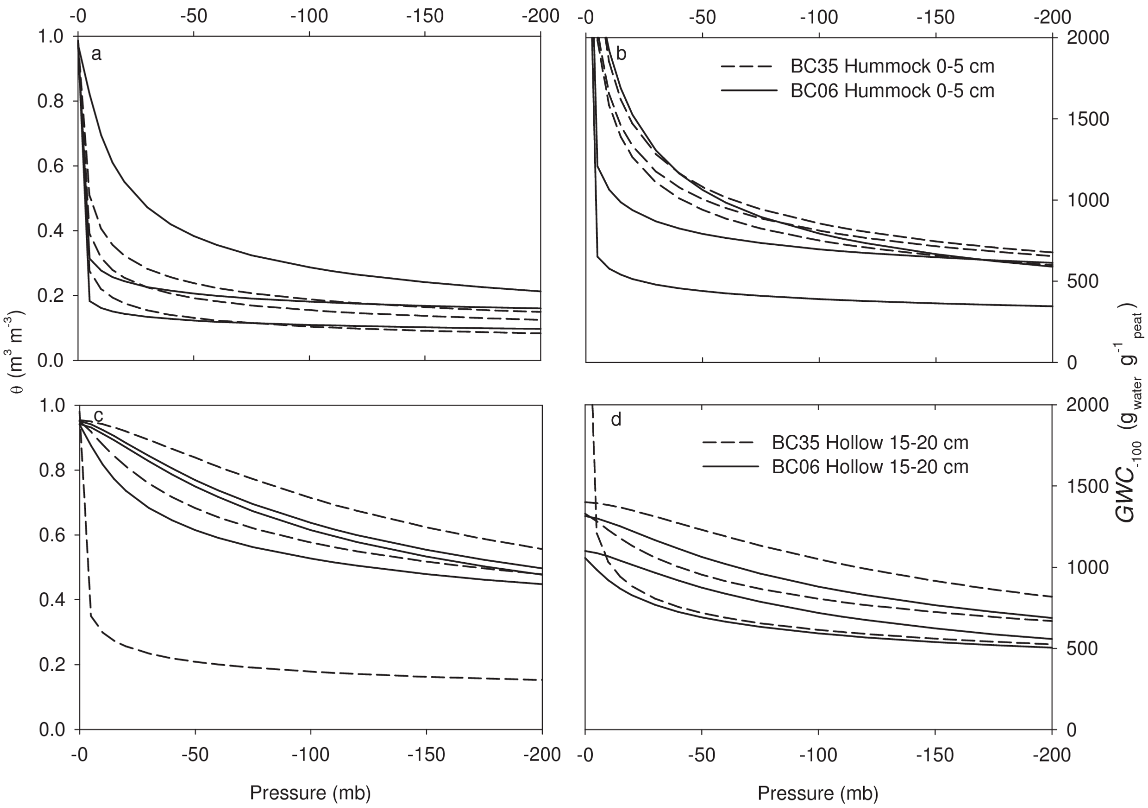

Figure 4.7. Water retention curves for peat samples from (panel a) the surface of hollows and (c) 15 cm depth in hollows. The equivalent water retention curve expressed in terms of gravimetric water content (*GWC*) is expressed in the right panels (b) and (d). Triplicate lines shown are water retention curves of three individual samples, not replicate runs.

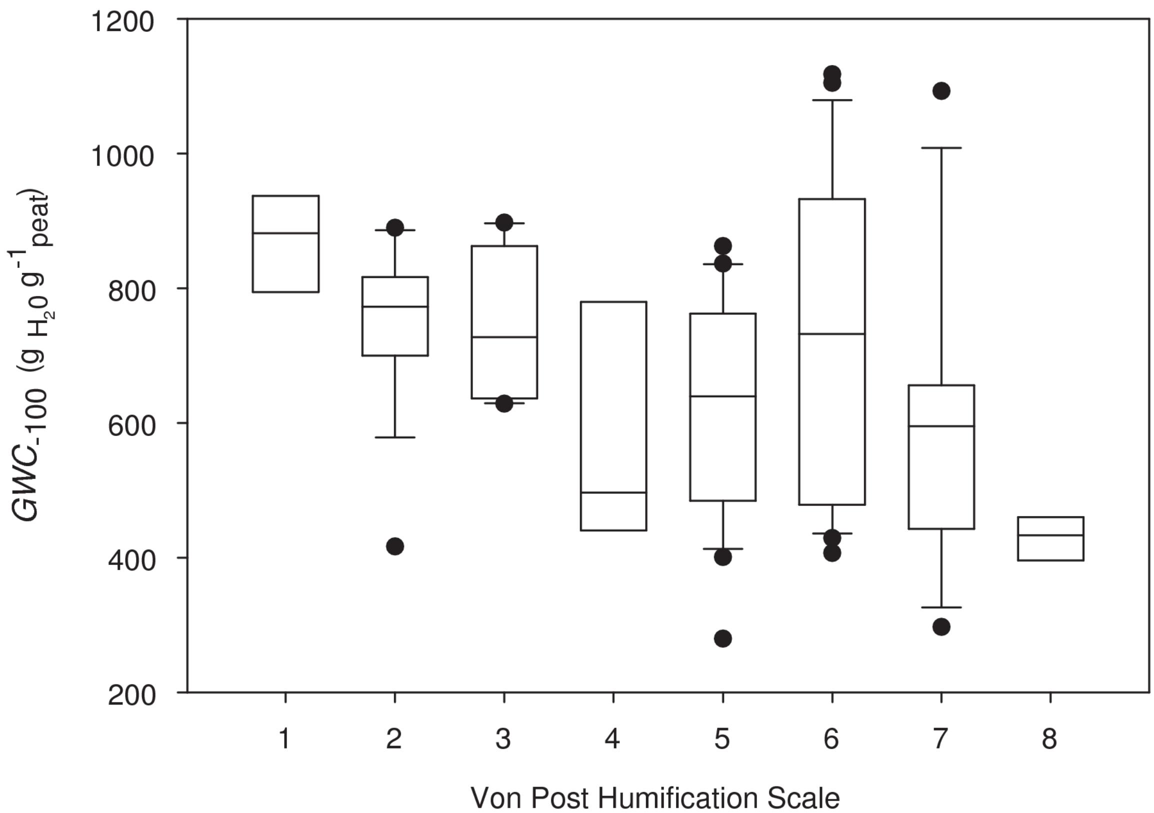

Figure 4.8. GWC_{-100} as a function of Von Post's humification scale.

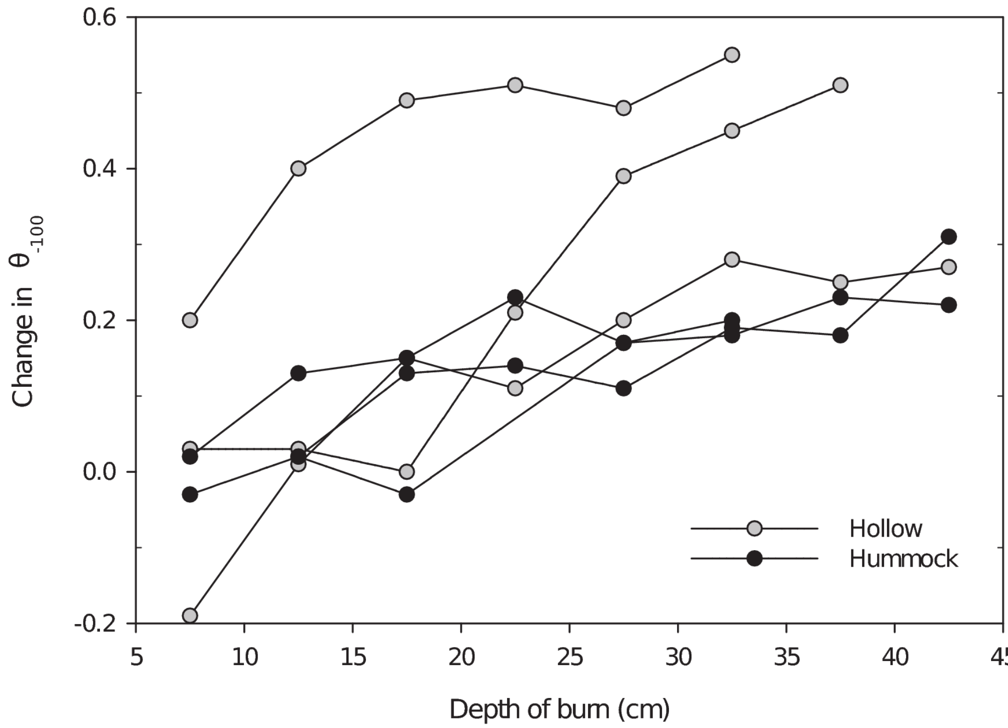

Figure 4.9. Modelled change in surface θ_{-100} as a function of depth of burn in both hummocks and hollows. Lines shown are for individual cores.

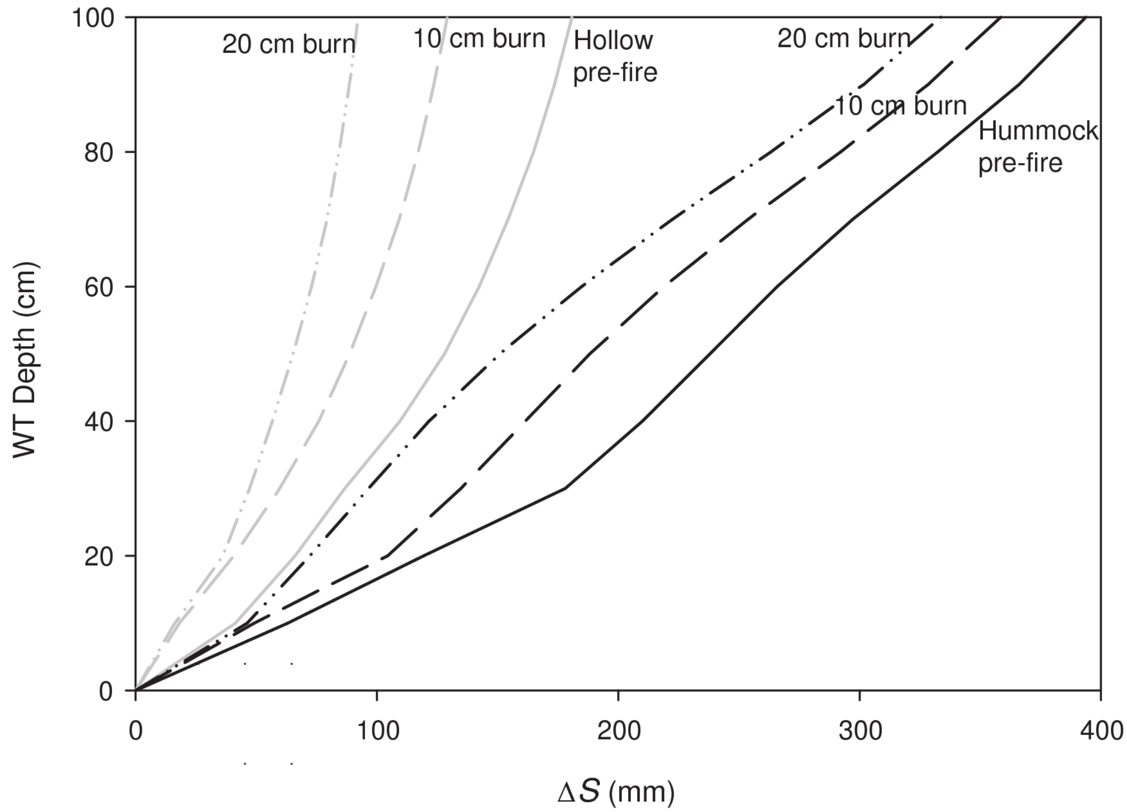

Figure 4.10. Water storage loss (ΔS, mm) required to achieve a water table decline in cm, for both hummocks and hollows (solid line). The effects of wildfire consuming surface peat to a depth of 10 and 20 cm on this storage relationship are shown as dotted lines.

CHAPTER 5: THE EFFECT OF WILDFIRE ON PEATLAND VADOSE ZONE HYDROLOGY

5.1. Abstract

Peatland response to wildfire disturbance has been shown to vary as a function of microform and vadose zone hydrology, with low-lying hollow microforms most susceptible to deep combustion. We examined a paired burned-unburned peatland to examine how wildfire disturbance alters vadose zone processes. Water table response to rain events increased significantly after wildfire, resulting in a more variable unsaturated zone thickness that was more responsive to smaller rain events. Water storage losses in the vadose zone occurred primarily at depths greater than 15 cm, with the volumetric water content at the surface far less variable than at depth. Significant water loss at the peat surface occurred in hummocks microforms in the early spring due to the presence of unsaturated frozen peat underneath. During such intervals, loss of storage in the vadose zone satisfied up to 25% of daily evaporative demand, compared to only approximately 3-5% during ice-free periods. Similar but less severe drying conditions were observed later in the summer, with burned hummocks being the most vulnerable to drying and high pore-water pressures. The enhanced surface drying observed is a precursor to high pore-water pressure conditions that can inhibit *Sphagnum* regeneration. Our observations point to a paradox where the hummocks, being most resistant to combustion, are themselves most prone to high pore-water pressures after fire. This more harsh environment in the hummocks may be a contributor to the observed delay in post-fire succession in hummocks compared to hollow *Sphagnum* species.

5.2. Introduction

Peatlands represent a globally significant carbon stock that is sensitive to hydrological conditions in both the accumulation (*e.g*. Tuitilla *et al*., 2004) and decomposition (*e.g*. Beer and Blodau, 2007; Benscoter *et al*., 2011). While some simple modelling studies suggest that peatlands display low resilience to disturbance such as climate change (Ise *et al*., 2008), other studies suggest that by accounting for the numerous negative feedbacks such as the water table-peat deformation feedback (Whittington and Price, 2006), peatlands reveal a high degree of resilience from enhanced water deficits. Negative feedbacks allow peatlands to exist at the margins of their optimal climate envelope (Yu *et al*., 2009; Halsey *et al*., 1995). However, peatlands in less than optimal climates, such as continental peatlands in the southern boreal forest, lack both the high precip rates of oceanic peatlands (Price, 1992) and the decreased subsurface flow of permafrost peatlands (Wright *et al*., 2009), and must relay on internal processes to minimize the frequency of substantial summertime water deficits. Forested ombrotrophic peatlands in the southern boreal forest are particularly vulnerable, as trees have the capacity to further reduce water availability at the surface under water table drawdown conditions (Rothwell *et al*., 1996). These forested peatlands are at additional risk due to wildfire disturbance, which kills the tree canopy and dramatically alters the vegetation community (Benscoter and Vitt, 2008). A key component to peatland resilience from wildfire disturbance is a cover of *Sphagnum* moss (Shetler *et al*., 2008), therefore the water relations and drying of *Sphagnum* is important for regulating the risk of deep smouldering and endangerment of the carbon pools within the peatland.

While the hydrological constraints on post-fire moss colonization are not fully understood, evidence from peatlands drained for horticultural peat extraction show a sensitivity to pore-water pressure at the surface, where *Sphagnum* recolonizing a bare peat surface has been found to be successful only in places where the pore-water pressure (Ψ) did not exceed (more negative than) -100 mb at any point in the summer (Price and Whitehead, 2001). This pressure corresponds to the air-entry pressure of the hyaline cells in *Sphagnum* mosses (Lewis, 1988; Hayward and Clymo 1982) which are adjacent to the photosynthesizing cells. While the post-fire vegetation succession shows a topographic contrast, where hollows are recolonized faster than hummocks (Benscoter, 2006), there is a lack of understanding of the hydrological conditions that *Sphagnum* undergoes in the post-fire environment. The distribution of pore-water pressure after fire in peatlands remains unknown, despite being a much larger disturbance by area (Turetsky *et al.*, 2004) compared to the more well-studied horticultural extraction disturbance. The aims of this study are to determine the pore-water pressure regime of peatlands before and after wildfire in order to determine if vadose zone conditions are a potential inhibitor of *Sphagnum* regeneration.

In unconfined aquifers such as peatlands, the water table (*WT*) provides a lower hydrological boundary condition of the vadose zone, where pore-water pressure and elevation head (z) are both equal to zero (Freeze and Cherry, 1979). Under the simple scenario of static equilibrium, the pore-water pressure (a negative value in mb) at any given depth in the unsaturated zone is equal to the depth to *WT* (equivalent to elevation head; Gilham, 1984). In unconfined aquifers such as peatlands, the uniformly high permeability of the aquifer precludes a perched aquifer within the peat, so pore-water pressure cannot be less than the distance to the *WT* except during the infiltration of rain (Lindholm and Markkula, 1984). In most circumstances, the vertical unsaturated hydraulic conductivity through capillarity is sufficient to conduct water from the water table to the evaporating surface (Päivänen, 1973). In such cases, the water table can be seen as being "connected" to the surface, and the pore-water pressure and water table depth are tightly coupled (Lindholm and Markkula, 1984). The water table has additional importance in predicting the depth of the oxic zone, an important control over tree rooting depth (Lieffers and Rothwell, 1987) and decomposition processes (Beer and Blodau, 2007; Frolking *et al*., 2001). As such, the effectiveness of rainfall in raising the water table is of interest, as it lowers the distance for vertical unsaturated flow to reach the surface. Nevertheless, the volumetric water content (θ) profile is a function of the depth to water table in peatlands (Hayward and Clymo, 1982). The θ profile in the unsaturated zone of a peatland in equilibrium can be conceptualized as a non-linear function from saturation at the water table, with little change in θ near the surface (Granberg *et al*., 1997). The capillary fringe may be included, which introduces a large increase in θ just above the water table (Price and Whitehead, 2004) that in a uniform substrate would approximate the water retention curve. Instances where the water table is hydrologically disconnected from the surface (here called disequilibrium) are typified by low θ in the upper 0-10 cm, followed by an abrupt increase in θ to those similar to equilibrium at depth (Price, 1997). Under disequilibrium conditions the Ψ profile also shows large negative values near the surface (Price, 1997) far in excess of the distance to the water table. Since the unsaturated hydraulic conductivity of peat declines exponentially with greater Ψ (Price and Whittington, 2008), drying of the surface in non-equilibrium conditions could lead to the development of a very dry peat surface with an unsaturated hydraulic conductivity many orders of magnitude lower than the soil immediate below. This dry surface effectively increases the surface resistance to evaporation (van deGriend and Owe, 1994; Kettridge *et al*., submitted), potentially providing primarily an evaporation-limiting cap. These disequilibrium conditions are found in peatlands only affected by drying impacts such as ditching, peat harvesting, or afforestation (Price, 1997; Lindholm and Markkula, 1984). Changes in surface θ after disturbance may be the result of disequilibrium (Price, 1997) or *in situ* alteration of the water retention properties of the peat through compression (Schlotzhaur and Price, 1999), or both. In such cases, a knowledge of the water table alone is insufficient, and additional measurements of surface Ψ or θ are required. A knowledge of the water content profile in peatlands is useful not only in understanding evaporation processes and moss growth limitations, but also in the study of peat combustion, where water content forms the energetic limitation to smouldering combustion (Benscoter *et al*., 2011). However, our current conceptualization of vadose zone processes in peatlands susceptible to wildfire barely extends beyond the simple equilibrium models such as Granberg *et al*. (1997).

In this work we apply a similar examination framework to that used in studies of cutover peatlands, but with wildfire as the disturbance of interest. In addition, the dynamics of water table variance will be examined as a lower boundary condition in the vadose zone. By combining water table dynamics with pore-water pressure measurements, we will examine hydrological disequilibrium in peatlands as a vulnerability assessment tool (*c.f.* Waddington *et al*., 2012). Lastly, water content measurements will be examined in the context of water storage in the vadose zone.

5.3. Site Description

Field studies were conducted in a northern Alberta peatland (55.81º N, 115.11º W) partially disturbed by wildfire in 2006. The remainder of the peatland was last disturbed by wildfire in approximately 1935 (hereafter referred to as BC35). The 2006 burn site (hereafter referred to as BC06) is part of a 44 ha wildfire from September 2006 that caused 100% tree mortality within the fire scar and burned over 99% of the peat area within the scar. Mean depth of peat consumption at BC06 was estimated to be 7.5 cm using the adventitious root method of Kasischke *et al*. (2008). The 1935 burn site (BC35) is located 350 m from the edge of the 2006 burn. The two observation plots were 800 m apart, allowing for identical atmospheric conditions.

These peatlands are typical of forested boreal, ombrotrophic peatlands of the continental high boreal wetland region (NWWG, 1986). The local abundance of peatlands on the landscape is 27% (Tarnocai *et al*., 2000). The long-term average (1971-2002) annual precipitation and temperature for Slave Lake is 502 mm and 1.7 ºC, respectively (Environment Canada, 2000) with an average annual open-water evaporation of 585 mm (Bothe and Abraham, 1993). Snowfall is a relatively small proportion of precipitation (29%), even though the snowpack is present for an average of 144 days per year (Environment Canada, 2000). The winter of 2010 was equal to the lowest recorded April 1st snowpack (0 mm), and 6th lowest March 1st (38 mm) in a 35 year record near Slave Lake, AB (Alberta Environment, 2010). As a result of the record-low snowpack, persistent seasonally frozen peat was observed at depth in hummocks until JD 220. In previous years, seasonally frozen peat was found to disappear on JD 180 or earlier.

Vegetation at both sites is dominated by *Sphagnum* mosses in a pattern of high areas 30-40 cm in elevation and spanning 1 m^2 or more (hummocks) and low features (hollows). Hummocks are largely composed of *Sphagnum fuscum* with a shrub canopy of *Ledum groenlandicum* that can reach up to 30 cm in height. *Rubus chamaemorus*, *Vaccinium vitis-idaea*, and *V. oxycoccus* grow immediately adjacent to the moss surface, and do not reach more than 5-10 cm in height. Hollows are a mix of *S. angustifolium*, *Cladonia spp.* lichens, *Pleurozium schreberi* feathermosses, and bare peat, with generally a lower density of shrubs. Trees at BC35 are exclusively *Picea mariana* with a mean height of 2.3 m at both sites and a basal area of 11.0 and 11.2 m^2 ha^{-1} at BC06 and BC35, respectively. The stem density is 19 700 and 16 050 stems ha^{-1} at BC06 and BC35, respectively.

Mean peat depth is 1.6 and 1.7 m at BC06 and BC35, respectively. The post-wildfire peatland environment is typified by largely unburned *Sphagnum* sheep (*sensu* D. Vitt), which are composed of *Sphagnum fuscum* hummocks, interspersed with more deeply burned hollows with a thick char layer. Combustion in hollows is nearly double that in hummocks (Benscoter and Wieder, 2003), likely due to differences in how moisture retention contrasts between these peatland microforms (Chapter 4). BC06 contained 62% hummocks by area, while 72% of BC35 was classified as hummocks.

5.4. Methods

5.4.1. Water Content and Storage

Peat volumetric water content (θ) was monitored at depths of 5 and 15 cm below the surface of hummocks and hollows using Campbell Scientific (Logan, UT) CS 616 time domain reflectometry (TDR) probes. Additional TDR probes were placed at 30 cm depths in the hummocks at both sites. Arrays of TDR probes were deployed at the same depth in both hummocks and hollows at both BC35 and BC06. Measurements were made every 60 minutes. T-type thermocouples were deployed alongside the TDR probes, or at an adjacent hummock or hollow at the same depths. The peat temperature was used to calculate the dielectric constant for water using Weast (1982). TDR period output was logged from the 616 probes and converted to a bulk dielectric permittivity using the equations from Hansson and Lundin (2008), and θ was calculated using the mixing model of Kellner and Lundin (2001). Because the probes were installed at a long-term monitoring site, peat bulk density at the TDR probes was not directly measured, but rather was estimated using peat cores taken from adjacent hummocks and hollows at each site. Ice has a dielectric permittivity of ~4, and is thus more similar to the air (1) than water (~78) in its propagation of TDR waves (Weast, 1982). The presence of ice in the peat matrix introduces a fourth component to a three component (air-water-peat) mixing models such as Kellner and Lundin (2001). While TDR calibrations including ice do exist (*e.g.* Nagare *et al*., 2011), we used the increase in apparent water content during thawing to approximate the timing and depth of ice in the peat, since the determination of the exact ice content is not the objective here.

Unsaturated zone water storage was calculated as the sum of storage at three layers in hummocks and two layers in hollows, with the midpoint of each layer representing a TDR probe (*c.f.* Yazaki *et al*., 2006). Each layer remained unsaturated during the period when θ were made, though ice persisted at some layers far into the summer of 2010. At both sites, the hollow layers are comprised of a 0-10 cm and a 10-22.5 cm layer, represented by the 5 cm and 15 cm probes, respectively. At the hummocks, θ was monitored to a depth roughly corresponding to the 15 cm hollow depth. At both BC35 and BC06, the upper horizons are 0-10 cm and 10-22.5 cm, with the BC35 hummock extending to a depth of 45 cm, while the BC06 hummock was only extended to a depth of 40 cm due to its lower vertical position compared to the adjacent hollow. End of day observations of θ were used to calculated the daily vadose zone storage in hummocks and hollows ($S_{\theta\mathrm{hum}}$, $S_{\theta\mathrm{hol}}$) as well as a site average vadose storage term (S_θ), given as a vertical depth of water in mm:

$$S_\theta = S_{\theta_{hum}} + S_{\theta_{hol}} = f_{hum} \sum_{i=1}^{n} \theta_i \Delta z_i + f_{hol} \sum_{j=1}^{n} \theta_j \Delta z_j \qquad (5.1)$$

where i and j represent depth intervals in hummocks and hollows and f represents the proportion of the microtopographic form on the landscape.

5.4.2. Micrometeorological and Water Table Measurements

Precipitation was logged at 20-minute intervals using a Texas Electronics (Dallas, TX) tipping bucket rain gauge accurate to 0.1 mm. Temperature and vapour pressure were measured at 20-minute intervals using a CS215 sensor (Campbell Scientific, Logan, UT, USA). Water table was measured at 20-minute intervals at each site using a Dataflow Systems Odyssey (Christchurch, NZ) capacitance water table logger or an Ott (Kempten, Germany) PLS pressure transducer.

5.4.3. Pore-water Pressure

Pore-water pressure was measured at five hummocks and five hollows at each site, with an identical depth series of 5 and 15 cm. Tensiometers at 30 cm depth were installed at the hummocks only. Soil Measurement Systems (Tucson, AZ, USA) tensiometers were used, and measurements of pore-water pressure were made with a UMS (Munich, Germany) Infield tensimeter, accurate to ±2 mb. Manual measurements were made weekly between 1200 and 1700h, and no less than 24 hours since the last significant rainfall. All tensiometers, wells, and CS 616 probes were surveyed with a Lecia Geosystems (Heerbrugg, Switzerland) Total Station to a common elevation datum at each site.

Under equilibrium conditions where the hydraulic head ($h = z + \Psi$) gradient is uniform throughout the unsaturated zone and assuming a conservative K_{unsat} of 120 mm d^{-1} (Price *et al*., 2008), a daily evaporative demand of 3 mm would only induce a hydraulic gradient of 0.025. In contrast, a 10 cm difference between z and Ψ even at the top of a high hummock under low WT (z = 100 cm) would induce a hydraulic gradient of 0.1, with shallower WT under the same pore-water pressure inducing larger gradients in turn.

As such, measured pore-water pressure was compared to the local depth to *WT* from the tensiometer cup, and pore-water pressure deviation ($z - \Psi$) of less than 10 cm was considered to be in equilibrium, where pore-water pressure is equal to depth to *WT* (Gilham, 1984). A threshold of 10 cm was used, as it is beyond expected error in both the tensiometer and elevation survey.

The occurrence of disequilibrium conditions was modelled using a logistic regression of tensiometer characteristics (hummock/hollow, site, tensiometer depth, absolute elevation) and environmental parameters (Julian Day, depth to *WT*, vapour pressure deficit). In order to account for daily variation in evaporative demand altering Ψ and disequilibrium, daily average vapour pressure deficit (D_z) was calculated using the method of Oren *et al.* (1996) and utilized in the logistic regression. The regression was performed using the *glm* package in the R statistical program version (R Core Development Team, 2011).

5.5. Results

5.5.1. Water Table Response to Precipitation

Rainfall events followed a power-law distribution, with 15% of the rainfall events representing 50% of the total summer rainfall in the summers 2009 and 2010. At BC35, 54% of all rainfall events resulted in no increase in water table position, while the same was true at BC06 in 42% of all rainfall events. The average amount of rain in events resulting in no positive *WT* response was 1.1 and 1.2 mm at BC35 and BC06, respectively. The rainfall threshold to invoke a WT response is slightly lower than the x-intercept of precipitation versus *WT* response (Figure 5.2a), which are 2.0 and 1.6 mm at BC35 and BC06, respectively. All rainfall events greater than 2.0 mm resulted in a net *WT* increase regardless of antecedent *WT* depth at BC06.

Mean water table response (mm water table change per mm of rain) across all rainfall events at BC35 was estimated by the slope of the linear regression in Figure 2a, and was equal to 3.2 and 5.7 at BC35 and BC06, respectively. The two slopes are significantly different at a 95% confidence interval ($t_{1,137}$ = 2.4, P = 0.016). A comparison of *WT* response for the same rain event (Figure 5.2b) shows that the magnitude of *WT* response at BC06 is on average 63% greater compared to BC35. Antecedent water table depth showed no effect on water table response at either BC06 (R^2 = 0.025; F = 0.456; P = 0.51) or BC35 (R^2 = 0.001; F = 0.008; P = 0.93).

5.5.2. Vadose Zone Water Storage

Given that vadose water storage changes with respect to *WT* depth (Figure 5.3), $S_{\theta\mathrm{hum}}$ and $S_{\theta\mathrm{hol}}$ was analyzed at a constant *WT* depth of -20 cm below hollow surface in order to highlight differences amongst sites, microtopography, and depth. Overall S_θ was higher at BC06 compared to BC35, where an areal average of 175 mm at BC06 compared to BC35 where only 80 mm was stored. $S_{\theta\mathrm{hol}}$ was similar in both sites at between 54 and 58 mm. $S_{\theta\mathrm{hum}}$, however, showed a large contrast between sites, with the hummock at BC35 storing only 90 mm while at BC06 220 mm was stored. While storage in the upper 10 cm of hummocks was slightly larger at BC06 (19 vs. 13 mm), observations at 15 cm depth revealed a doubling of storage at BC06, and three times the storage was observed at the 30 cm depth probes at BC06 compared to BC35.

The change in S_θ, rather than absolute values, is a function of the retention properties of the peat, the *WT* (as a lower boundary condition), and any excess evaporation at the surface which leads to disequilibrium conditions. The summer of 2009 was relatively free of seasonally frozen peat at depth from JD 180 onwards, and was subsequently used to examine ice-free conditions. The absolute amount of water lost in the vadose zone of hummocks at BC35 and BC06 was similar at 21 and 21.5 mm (Figure 3), and can be considered identical given the errors in measurement. *WT* drawdown from the net water deficit in 2009 resulted in a 22% decline in water storage in hummocks under a *WT* falling from 43 to 66 cm below the hummock surface. The majority of water lost in the hummock profile was recorded at 30 cm depth (a 30% decline in storage). The upper 10 cm saw a decline of only 8% (from a θ of 0.13 to 0.12), even though it represented 22.5% of the total vertical extent of the profile. The lowest layer, from 22.5 to 45 cm below the surface, was responsible for 82% of the water volume lost, despite being 50% of the total vertical extent. At BC06, the hummock experienced a smaller decline of 10% in profile storage, despite a larger *WT* decline from 38 to 70 cm below the hummock surface (Figure 5.1). The extent of drying was more uniform in the BC06 hummock, with storage losses ranging between 8-14% and the largest proportional loss of storage occurring in the upper 10 cm, though the magnitude of the loss (3 mm or θ of 0.03) was small. While each of the three layers monitored in the hummock at BC06 lost similar proportions of the water stored in spring, the magnitude of losses increased from

3-12 mm, increasing with depth. The three horizons contributed to the total water lost in the profile roughly proportional to their vertical extent, 55% (12 mm) of the total storage loss originating from the lowest horizon, which occupies 44% of the vertical extent.

The hollows, despite being more modified by wildfire, showed a smaller contrast in vadose zone dynamics. Absolute storage loss in the hollow at BC35 was larger in 2009 at 12.5 mm, while the BC06 hollow showed a smaller loss of 9 mm. Both the charred upper 10 cm of the hollow at BC06 and the *Sphagnum angustifolium* hollow at BC35 showed between 15-16% storage loss during the summer. The lower hollow horizon at BC35 showed a larger proportion of storage loss at 28%, compared to only 16% loss at BC06, despite a consistently lower *WT* at BC06. The proportion of water loss was similar at both sites, with the upper (0-10 cm) horizon at both sites responsible for 40% of the total storage loss, while occupying 67% of the total vertical extent of the profile.

Overall, vadose zone water losses in hummocks were equivalent to 10% of the cumulative surface evapotranspiration during the study period in 2009. In hollows, the proportion was smaller at 6 and 4% at BC35 and BC06, respectively. During short rain-free intervals, the changes in vadose zone water storage were a larger proportion of evapotranspiration. Between JD 201 and 207 in 2009, vadose storage losses in the BC06 hummock averaged 1.1 mm d^{-1} (Figure 3), while daily evaporation was 4.5 mm, meaning 25% of daily evaporative demand was satisfied by storage loss in the vadose zone. Similar declines in unsaturated storage were observed later in the summer of 2009, though the rate of decline in storage decreased as the water table elevation itself fell.

5.5.3. Effect of Seasonally-frozen Peat

Record low snowfalls during the winter of 2009-2010 provided an opportunity to observe the effects of seasonal near-surface ice on the hydrology of the overlying unfrozen peat. Rainfall in the summer and autumn of 2009 was low, resulting in a *WT* depth of 70 cm below the peat surface at freeze-up. Snowfall was limited to 28 mm snow water equivalent during the entire winter (Environment Canada, 2011), and albedo measurements at BC06 (Chapter 1) show that snow cover was less than 50% by JD 69. Remote sensing of snow water equivalent (Environment Canada, 2011) revealed the snowmelt was complete before JD 90. Rainfall events on JD 109-113 thawed the upper 10 cm of the hummock at BC06 rapidly, while the hummock remained frozen to a depth greater than 30 cm. Between JD 114-137, only 3.7 mm of rain fell while potential evaporation totalled 70 mm. Average air temperature was 6.3°C, and daily maxima reached up to 26°C. As a result of the water deficit, θ at 5 cm depth in the hummock declined from 0.18 to under 0.12, while observed θ at 15 cm depth increased from 0.17 to 0.42 as the lower depth of the hummock thawed (Figure 5.4). During the same period, total loss in vadose storage in the upper 10 cm of the hummock was 6 mm. Unsaturated water storage loss from the upper 10 cm of the hummock satisfied up to 25% of the evaporative demand during this interval. In contrast, during ice-free periods unsaturated storage losses in the upper 10 cm represented only 2-3% of daily evaporation. Given an ice content of approximately 15%, with a daily decline in the frost table of approximately 3 mm, only 0.4 mm of water would be yielded daily from the melting of ice, while mean daily potential evaporation was 3 mm.

Using the relationship between *WT* depth at θ at 5 cm during ice-free periods, an equivalent *WT* depth (or pore-water pressure) was calculated down to 82 cm just prior to rainfall on JD 138 (Figure 4). In contrast, manual observations of *WT* at the site were only 36 cm below the hummock surface on JD 136, indicating the presence of unsaturated ice below the hummock in early spring significantly decreased θ in excess of equilibrium *WT* conditions. In fact, the θ of 0.12 observed on JD 136 was the lowest recorded θ during the summer of 2010, despite *WT* depths of over 55 cm below the hummock surface during latter parts of the summer. Similar short-term drying intervals (only 1.2 mm of rain over 16 days) only produced a decline in θ of 0.02 during periods when ice was absent from the hummock. A similar springtime decline in θ was observed in the hummock at BC35. Declines in θ of 0.07 were also observed in the 5 cm depth measurements in hollows at both sites from JD 114-137, but the resultant θ was in equilibrium with the observed *WT* measured on JD 136.

5.5.4. Pore-water Pressure

The distribution of pore-water pressure (Ψ) across all sites conformed well to a log-normal distribution ($\mu = 3.64$, $\sigma = 0.63$) once expressed as a positive pressure (also in mb) rather than as a negative. A Tukey's HSD test revealed the distribution of the natural logarithm-transformed values were not significantly different between hummocks and hollows at BC35 ($P = 0.15$). Moreover, Ψ values in hummocks at BC35 and hollows at BC06 were not significantly different ($P = 0.67$), though pore-water pressures were more negative at BC06 compared to BC35 ($P = 0.005$). In all but the hummocks at BC06, median Ψ was between -28 and -36 mb (Figure 5.5). In contrast, median Ψ in hummocks at BC06 was -62 mb. Similarly, the 10th percentile of Ψ was -55 mb in both hummocks and hollows at BC35, while the 10th percentile of Ψ at BC06 was -111 and -68 mb for hummocks and hollows, respectively. The effect of wildfire on Ψ was most apparent in hummocks, where the 10th percentile at BC35 (-55 mb) was only equal to the 66th percentile at BC06.

Observations of Ψ were found to be in disequilibrium at 5 cm depth 61% of the time in hummocks at BC06 compared to 36% at BC35. In hollows, disequilibrium was observed 14% and 34% of the time at BC06 and BC35, respectively. The logistic regression model applied to the risk of disequilibrium conditions was successfully applied (log likelihood = -227, d.f. = 11, $P < 0.001$). The regression model revealed that 5 cm depths were 13 times more likely to be at disequilibrium compared to 30 cm depth tensiometers, and that 15 cm tensiometers were three times more likely to be at disequilibrium compared to 30 cm depth tensiometers. The odds ratios from the logistic regression agreed well with the empirical probability distributions in Figure 5, with hummocks being nearly five times as likely to be at disequilibrium, while measurements at BC06 showed an odds ratio of 3.6.

Of the environmental variables measured alongside pore-water pressure, Julian day was the only variable to decrease the odds of disequilibrium, with a small but significant decline in the odds of 2% per day. Daily vapour pressure deficit (D_z) was the strongest determinant of disequilibrium, with a 1 kPa increase resulting in a 4.7-fold increase in the risk of disequilibrium (Figure 6). Local *WT* (depth from the tensiometer cup to the *WT*) was a strong determinant of disequilibrium, with each 1 cm of further distance from the *WT* resulting in a 7% increase in the odds of disequilibrium. Days since rain and the sum of rainfall in the previous 48 hours did not significantly alter the odds.

5.6 Discussion

5.6.1. Comparison to Cutover Peatlands

The θ values for surface peat shown here are considerably lower than those from cutover peatland ecosystems, where θ regularly exceeds 50% (Price and Whitehead, 2001). After remediation, median θ in recovering cutover peatlands can exceed 80%, in large part due to the high surface bulk density (Petrone *et al*., 2004). This is primarily due to an increase in bulk density at the surface due to the effects of harvesting, which removed the low-density peat to a far greater extent (80-100 cm) compared to wildfire disturbance (0-20cm). The trend of increasing vadose zone variability with depth is consistent with cutover peatlands, where θ is least variable at the surface, and increases with depth (Ketchison and Price, 2009). While we observed more variable water storage after wildfire compared to the unburned site, our dataset is not readily comparable to that of Ketchison and Price (2009). Compared to storage changes in undisturbed unforested peatlands (*e.g* Yazaki *et al.*, 2006), the losses observed here are larger at both the unburned and burned sites.

The resilience of the hydrological regime of the hollows is notable, particularly in light of the greater depth of burn and the visual impact of the surface char layer, which can create a bias towards assuming the hollows suffer a greater impact from fire. The hollows experienced much more rapid recolonization of species such as *Polytrichum strictum*, a precursor nursery species to *Sphagnum*, at BC06 within a year after fire. At the same time, the hummocks experienced slower recolonization, on pace with the observations of Benscoter (2006).

At BC35 and BC06, θ measurements point towards a height of the capillary fringe which is smaller than the 40 cm reported by Price and Whitehead (2004) for cutover peatlands in Québec, Canada. In cutover peatlands, upwards of 80 cm of peat is removed during the harvesting process (Girard *et al*., 2002), which exposes much denser peat to the surface compared to the post-fire peatlands in this study. Since the height of the capillary is directly related to pore-size (Gilham, 1984) and hence bulk density, one can expect a generally thinner capillary fringe in peatlands affected by wildfire compared to the cutover environment. Gilham (1984) estimated the height of the capillary fringe based on laboratory water retention curves, where the height of the capillary fringe was equal to the maximum pore-water pressure applied before θ dropped below saturation. Using this approach, the laboratory water retention tests in Chapter 4 suggest that the capillary fringe is likely less than 10 cm in height.

The observations of pore-water pressure in the BC06 hummock shown above are similar to those from a mulched cutover peatland surface in Quebec, Canada by Price (1997). At the same site as the cutover peatland environment studied by Price (1997), the median summertime pore-water pressure in an remediated site was -130 mb, or 80 mb greater than the pore-water pressure at a nearby remediated peatland and well beyond the established threshold for *Sphagnum* colonization of -100 mb (Price and Whitehead, 2001). In naturally regenerating hummocks of *Sphagnum* on the cutover surface, Ketchison (2011) found the pore-water pressure to be largely in equilibrium with the water table, with pore-water pressures far less than threshold values of -100 mb. However, it is difficult to distinguish whether hummocks only regenerate in areas in equilibrium, or whether the hummocks themselves promote equilibrium and a more stable hydrological regime. In a cutover Finnish peatland, Lindholm and Markkula (1984) observed both the hummocks and hollows to be largely in equilibrium during a wet year, but the hummocks showed considerable disequilibrium during a drier year, with pore-water pressures over -500 mbar with a water table of only 50 cm. Pore-water pressures observed by Lindholm and Marrkula (1984) showed considerable sensitivity to rain-free intervals, with sustained pore-water pressures over -100 mb after 10 rain-free days. This is in contrast to the observations here, where days since rain had no effect on the pore-water pressure. This could be due in part to the higher specific yield of the burned sites compared to drained peatlands, which leads to a smaller water table drawdown during rain-free intervals. Moreover, the post-fire hollow surface at BC06 shows evidence of charred surface crusts that may limit evaporation during dry intervals between rain events (Kettridge *et al*., submitted).

5.6.2. Comparison to Peatlands Drained for Forestry

The observed water table response to precipitation of 3.2 at BC35 is slightly higher but comparable to values near two at an oligotrophic spruce swamp and open bogs in Finland from Reinikainen *et al*. (1984). A high water table responses of over five observed at BC06 were higher than any ombrotrophic peatland observed in Reinikainen *et al*. (1984), though open fens with considerable connection to upland slopes and groundwater were shown to have storativities of between 4 and 7. Hillman (1997) found that forested peatlands with a 30-50 m ditch spacing in central Alberta experienced an increase of water table response between two to three times those prior to drainage. Similarly, Rothwell (1982) observed a 1-3 fold increase in water table response after drainage and logging in a mixwood forested wetland with a shallow (20-30 cm) organic layer underlain by an organic-rich silty clay soil.

5.6.3. Importance of Spring Hydrological Processes

The vadose zone hydrology of organic soils above frozen horizons is poorly constrained, despite many decades of research into the dynamics of permafrost features such as palsas (*e.g.* Zoltai and Tarnocai, 1971; Hinkel *et al*., 2001). Few field observations of peat θ above frozen organic soils exist in the literature. Yazaki *et al*. (2006) measured peat θ at 5 cm depth above frozen peat at 15 cm depth, but no decline in near-surface θ was observed. During ice-free periods, storage losses in the upper 10 cm constituted less than 1% of evaporation. This lack of drying could be attributed to the oceanic and low potential evaporation environment of northern Japan where the observations of Yazaki *et al*. (2006) took place. Recently, Nagare *et al.* (2011) proposed a conceptual model of a dry surface layer on top of a permafrost peat plateau (a hectare-scale feature) due to an underlying convex saturated ice-lens that promotes the lateral runoff of snowmelt and rainfall inputs in the spring. Moreover, Nagare *et al.* (2011) found that in lab experiments the upper surface layer experienced moisture redistribution in the spring from unfrozen to frozen areas of the peat due to vapour gradients, which further promoted the drying of the surface layers by effectively creating a moisture loss interface at both the top (evaporation) and bottom (vapour flux to frozen peat) of the unfrozen hummock surface. Our observations here depart from the lab experiments of Nagare *et al.* (2011) in that the hummocks at both BC06 and BC35 were unsaturated to a depth of over 40 cm, and thus the frozen peat layers below the thawed peat were unsaturated. However, capillary rise above a frozen, unsaturated soil horizon may be limited by the smaller surface tension of the ice-water interface of 0.03 N m^{-1} (Sutherland and Gaskin, 1973) compared to 0.07 N m^{-1} for a peat-water interface (Valat *et al*., 1991). Hydrophobicity of organic matter enduced by wildfire (*e.g.* Varela *et al*., 2007) would only serve to further decrease the surface tension and potentially capillary rise. Under such a scenario, the LaPlace equation of capillarity gives a height of capillary rise that is less than half that of an ice-free state. Presumably the vertical hydraulic conductivity is affected accordingly, but no studies to date have observed the contrast in K_{unsat} above a saturated peat horizon versus a frozen one. Unfrozen K_{unsat} in peat is typically high, on the order of 5 mm h^{-1} for poorly decomposed peat at Ψ of -100 mb (Price *et al*., 2008). Using a model of K_{unsat} for frozen mineral soils (Lundin, 1990), there is a three order of magnitude decrease in K_{unsat} when ice content increases from zero to 15%. Thus K_{unsat} at 15 cm below a hummock would only be 0.005 mm, which is well below the rate of potential evaporation.

The early-spring stress mechanism observed here, while only observed for a ~10 day period, occurs during the spring where prolonged (2 weeks plus) rain-free periods occur regularly, with such periods further associated with large wildfire events in Alberta

(Quintilio *et al*., 2001). The months of April and May show significantly lower mean daily vapour pressure in the Slave Lake region of Alberta compared to the rest of the summer (Environment Canada, 2000). While in more humid and oceanic peatlands this post-snowmelt period is associated with the highest observations of θ (Lafleur *et al*., 2005) and water table (Verry, 1997), the low snowfall and high snow ablation rates (Pomeroy *et al*., 2002) in the boreal plain ecozone limit the snowmelt recharge of the vadose zone in hummocks.

5.7. Conclusions

The increase in water table response to rain events after wildfire results in a more variable water table regime, with a magnitude of change lower compared to either forestry or drainage impacts. The observed increase in water table response was disproportionately large compared to the loss of rainfall interception in the tree canopy experienced after wildfire. At BC06, the greater response of the water table to rain events points to enhanced routing of precipitation directly to the water table. Greater vadose zone storage observed in the burned hummock was coupled with greater vadose zone losses during the summer, suggesting that any advantages in peat water retention due to post-fire compression are likely muted due to increased evaporation. Delineating the role of post-fire peat compression versus the combustion of surface peat layers in altering vadose zone storage dynamics was not possible due to the lack of pre-fire peat elevation measurements, but is worthy of future study.

After wildfire, the hydrological regime of hummocks underwent more change compared to hollows. This greater impact occurs despite a smaller depth of burn during wildfire, and *S. fuscum* and other hummock-forming species as keys to resilience as ecosystem engineers. The cumulative impact to the hummocks is manifested as enhanced peatland water storage losses, overall greater pore-water pressure, and a greater probability of hydrological disequilibrium. Our examination of the nature of hydrological disequilibrium in post-fire peatlands revealed three distinct time scales where hydrological processes can lead to disequilibrium conditions and potential stress for regenerating mosses: daily (vapour pressure deficits), weekly (late spring ice), and seasonal (*WT* decline). Examinations of post-disturbance vegetation succession in peatlands should take into account the varying time scales as well as atmospheric and thermal processes that contribute to hydrologically-induced moss stress.

5.8. Acknowledgements

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Table 5.1. Logistic regression model of hydrological disequilibrium in pore-water pressure.

Factor	Beta Est.	Odds Ratio	P(>|z|)
(Intercept)	-4169		0.049
JD	-0.0133	0.99	0.021
Hum/Hol	1.578	4.84	<0.001
burn/unburn	1.285	3.61	0.014
Days since rain	0.0256	1.02	0.545
Vapour pressure	1.553	4.72	<0.001
Elevation	6.401	602	0.049
Local WT	0.0748	1.07	0.018
48h Precip	-0.00147	0.99	0.885
-15 cm depth	1.205	3.33	0.086
-5 cm depth	2.581	13.2	0.007

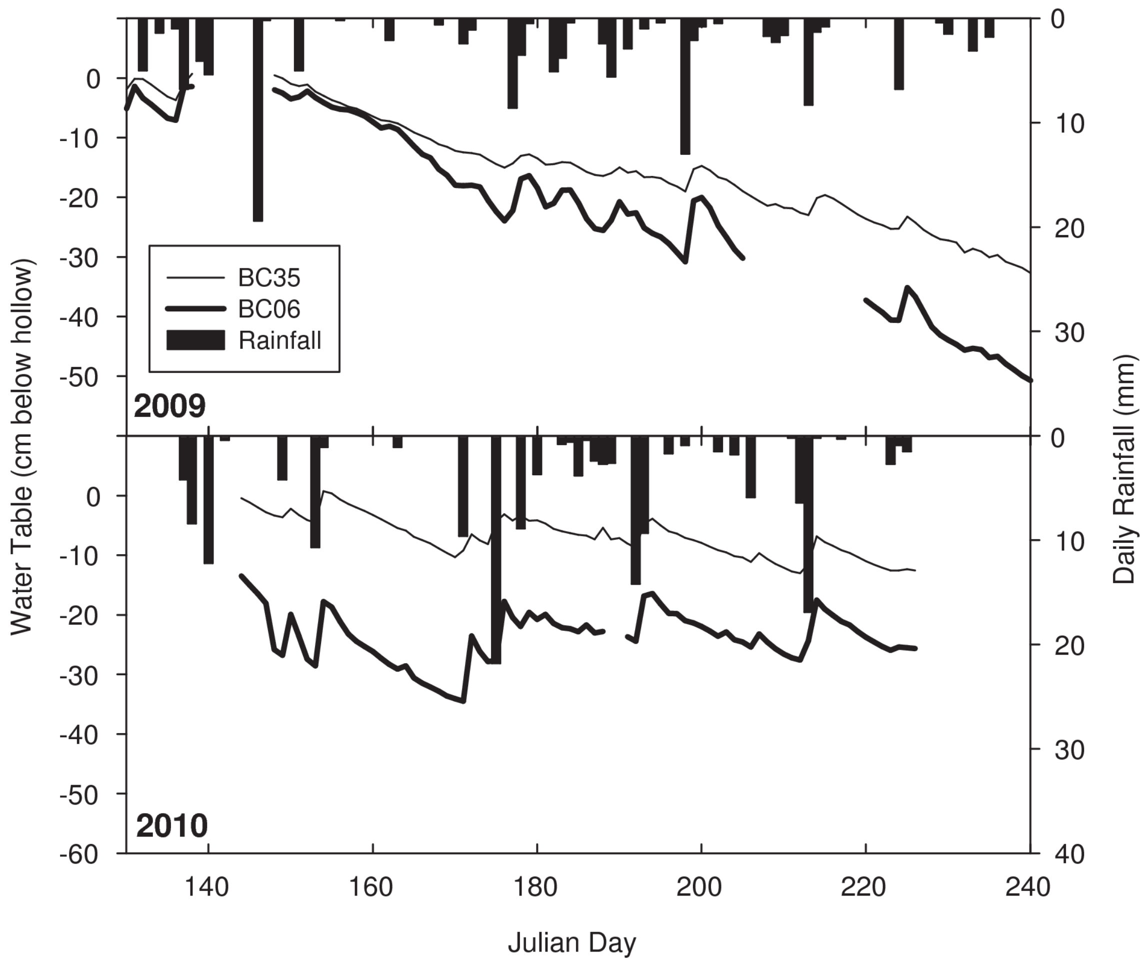

Figure 5.1. Daily rainfall and water table position at both sites during the study intervals in 2009 and 2010.

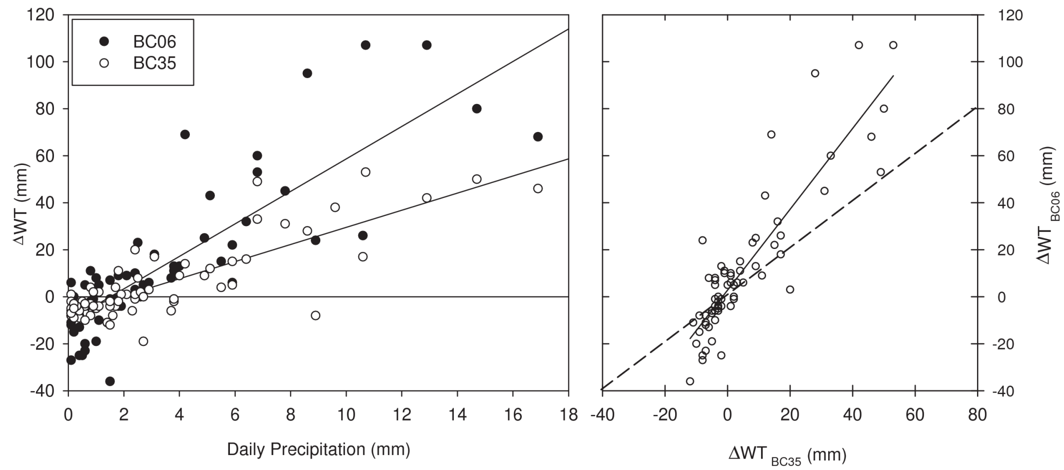

Figure 5.2. Water table elevation response (mm) to rainfall events in both 2009 and 2010 across both sites (panel a). Regression lines are shown for individual sites. Panel b is a comparison of water table response between both sites for the same rainfall event. The regression line is shown as a dashed line, while 1:1 relationship is shown as a solid line.

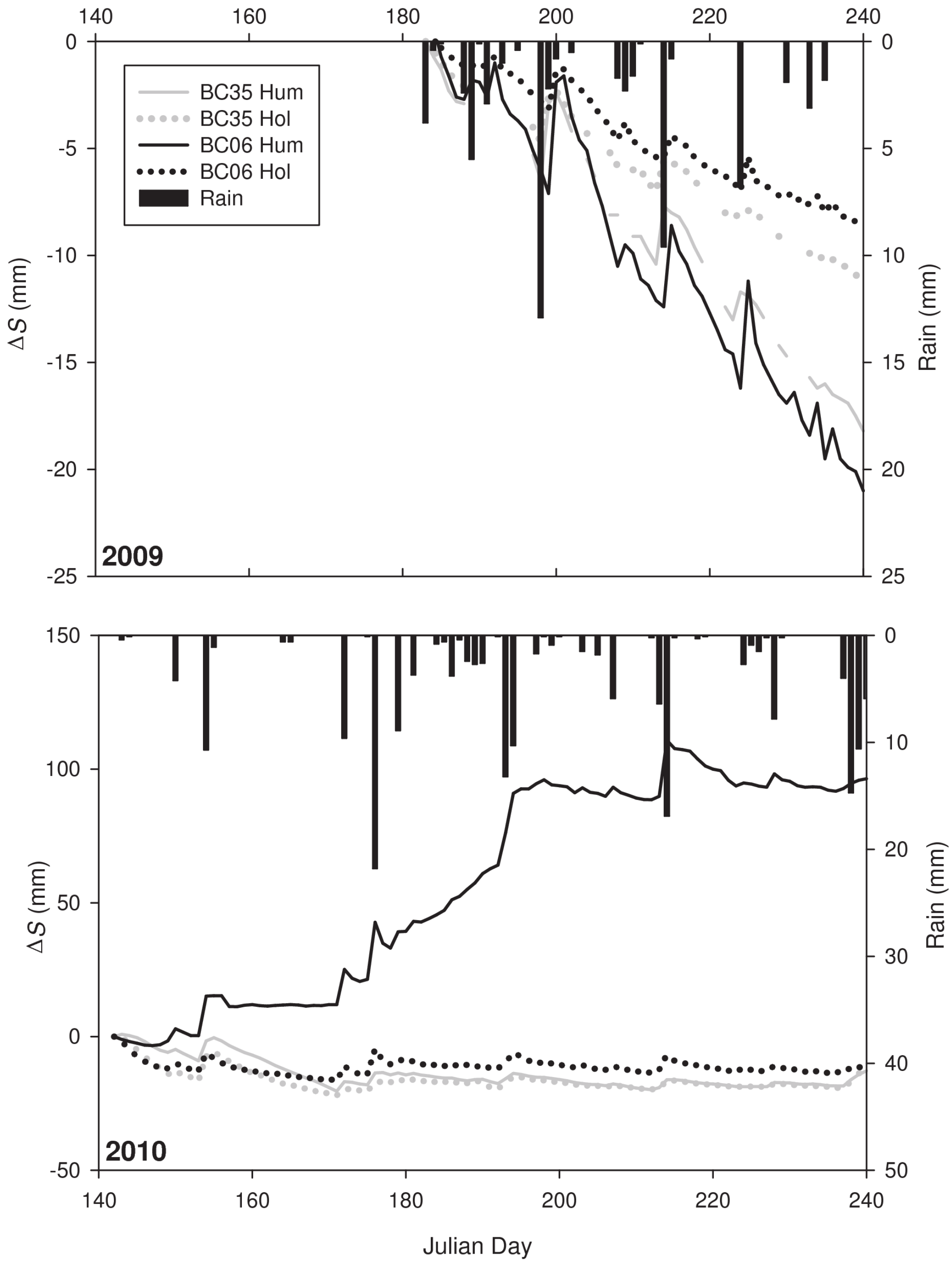

Figure 5.3. Change in vadose zone storage over time as a amongst hummocks and hollows and BC06 and BC35. Note that the vadose zone storage only accounts for liquid water stored, and the thawing of ice in the BC06 hummock during 2010 results in a net gain in storage.

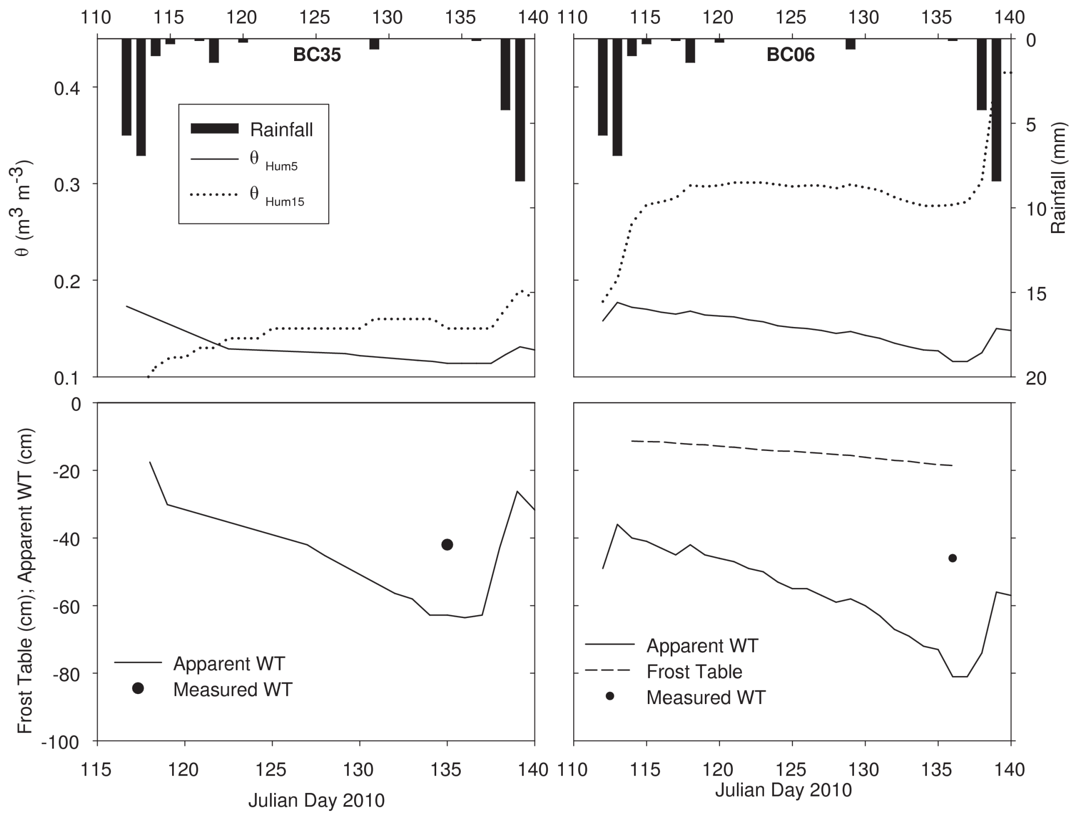

Figure 5.4. Impact of near-surface frozen peat on unsaturated zone hydrology of hummocks. Volumetric water content of measured at 5 and 15 cm depths at BC35 (a) and BC06 (b). The corresponding estimated pore-water pressure ("apparent WT"), and measured water table are given in panels (c) and (d) for BC35 and BC06, respectively.

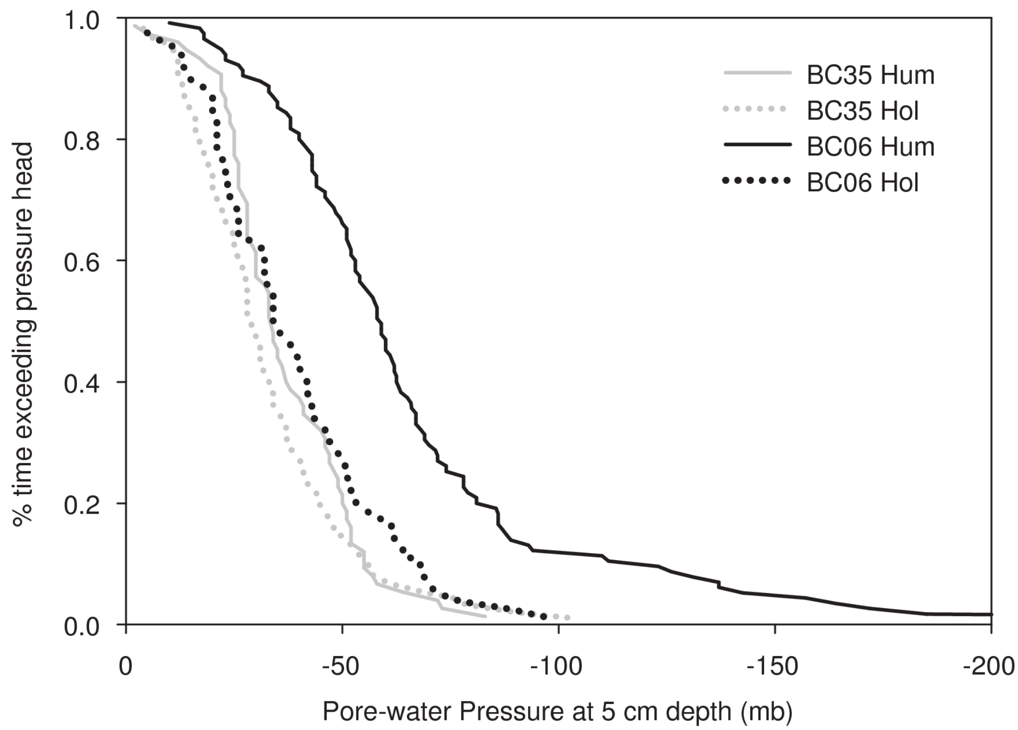

Figure 5.5. Distribution of pore-water pressure (Ψ) measurements at 5 cm depth from hummocks and hollows at BC35 and BC06.

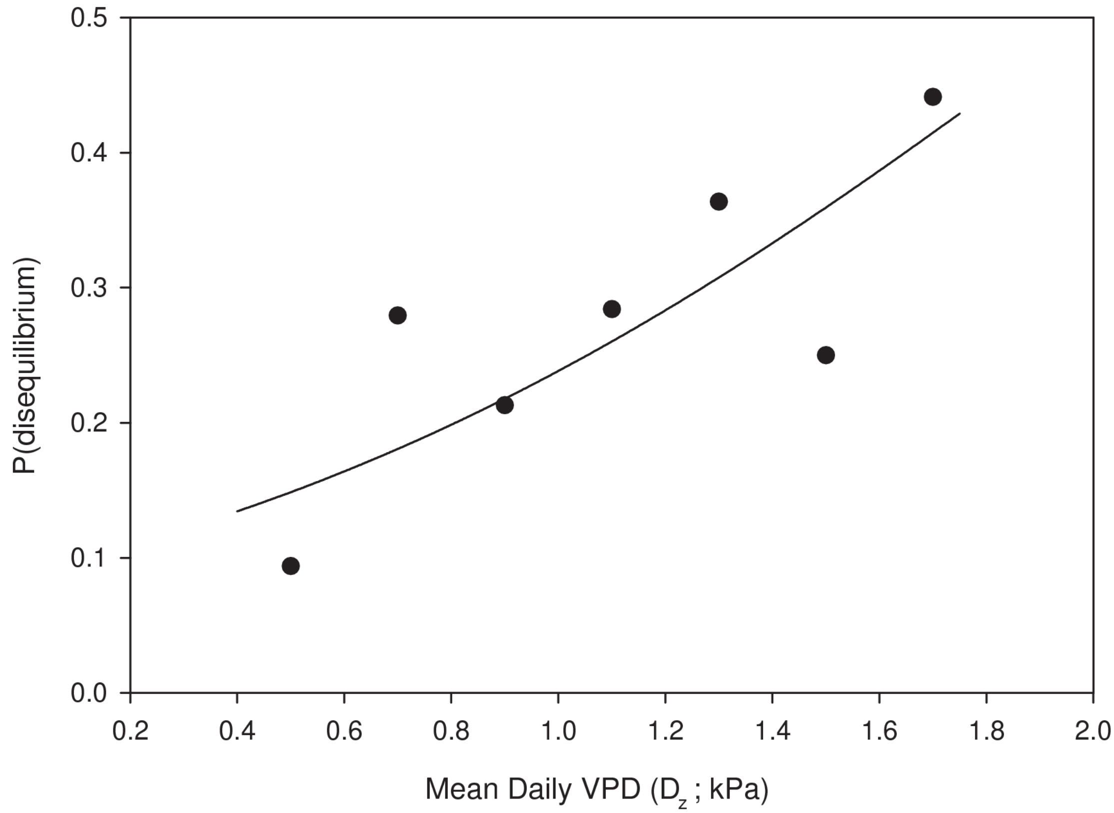

Figure 5.6. Probability of hydrological disequilibrium as a function of mean daily vapour pressure deficit. The data are presented as bin averages with 0.2 kPa bins. The fitted logistic regression is shown as the curved solid line.

CHAPTER 6: SUMMARY AND CONCLUSIONS

6.1. Summary

The impacts of wildfires on peatland ecohydrology were examined with an emphasis on the changing role of vegetation communities in environmental processes such as solar radiation exchange and water cycling. Albedo formed one of the most apparent changes to a landscape after fire. While albedo measurements after wildfire in forested peatlands was initially quite low, summertime albedo quickly rebounded within two years to levels similar to those prior to the fire. The observed recovery of the snow-free albedo was driven largely by a rapidly regenerating shrub layer, which muted the importance of the otherwise blackened hollow char surface of hollows. Winter albedo was largely driven by changes in the tree canopy, with the burned peatland experiencing large winter increases in albedo and a significant delay in the initiation of a daily positive net radiation flux to the surface.

The increase in canopy openness after fire enhanced surface longwave radiation losses at night. Enhanced night time radiative cooling resulted in the potential for enhanced dewfall and frost action on the peat surface. Coincident with the increased night time longwave losses was increased solar radiation fluxes to the peat surface following fire, which resulted in an overall increase in energy available for evaporation. Moreover, the spatial variance in the surface radiation budget decreased markedly after fire, with heavily shaded areas in the vicinity of tree clusters alongside open *Sphagnum* lawns and hummocks transformed into a more uniform radiation regime dominated by moderate shrub shading.

Given the large increase in energy available for evaporation, actual evaporation increased greatly at the surface following fire. Moisture availability was generally not significant in limiting evaporation on the landscape, as only a minor and not significant increase in surface resistance to evaporation was observed. Increased evaporation was offset by a complete halt to transpiration after fire, though the contribution of transpiration prior to fire was minor compared to surface evaporation. Overall, transpiration was far less in the open-canopy black spruce stands compared to a closed-canopy forest on mineral soils. The role of the sparse canopy in reducing incoming shortwave radiation outweighed the transpiration of the trees themselves, suggesting that trees in peatlands may serve as water conservation agents under ideal circumstances of moderate tree density.

Despite the lower water table in drier years, the vertical depth of peat compression was similar between burned and unburned sites, meaning the peat was less compressible after wildfire. This observation coincides well with an increase in measured bulk density near the surface after fire, largely through the preferential combustion of less dense peat at the surface. Additionally, there is the possibility of significant immediate post-fire peat compression that was not recorded, but would be analogous to that observed in drained peatlands.

A major outcome of the increase in bulk density was the increase in the water retention capacity of the peat near to the surface. The increase in water retention came at the expense of the volume of water available for use by *Sphagnum* mosses, since a smaller volume loss was required to reach pressures corresponding to moss stress in more dense peat. There was some evidence of a loss of water retention with no coincident change in bulk density in hummocks that showed little char formation. However, a change in water retention after fire was noted in peat horizons with significant coarse and fine root content, which transitioned from a high water content and active water transport medium (live roots) to a dry and low density component of the peat matrix after fire and the death of the above-ground vegetation. The humification state of the peat was found to interact with bulk density in reducing water retention per unit dry weight, such that very humified peat was less able to retain water compared to moderately humified peat of the same density.

The net effect of lower water tables, higher surface evaporation, and higher water retention was to simultaneously decrease volumetric water content at the surface and induce greater pore-water pressures observed after wildfire. Overall, hummocks experienced the greatest increase in pore-water pressure after fire, resulting in a surface hydrological regime more hostile to *Sphagnum* regeneration, despite the greater depth of burn in the hollows. The unsaturated zone transmitted water from the water table to the atmosphere at equilibrium rates a majority of the time. However, after fire a three-fold increase in the occurrence of hydrological disequilibrium (preferential drying at the surface) was noted after fire. The frequency of this disequilibrium was greatest in hummocks after fire. Enhanced pore-water pressure in post-fire hummocks was the result of disequilibrium conditions, and rarely the result of deep water tables alone. The occurrence of disequilibrium conditions in early spring was notable for the surface water contents recorded at hummocks in both the burned and unburned site. Deep, unsaturated ice present in hummocks limited capillary rise from the saturated zone. Coupled with long days with low vapour pressure, the surface peat was effectively drying at two fronts: to the atmosphere and to the lower vapour pressure in the frozen peat underneath.

6.2. Significance

The changes observed in individual ecohydrological processes of forested peatlands after wildfire disturbance are substantial, but do not appear to be sufficient to cause a wholesale shift from the ombrotrophic peatland ecosystem observed prior to wildfire. Ericaceous shrubs and the herbaceous layer appear to contribute significantly to the resilience of a peatland to wildfire disturbance. While the role of shrubs has been examined in shrub-dominated peatlands such as Mer Bleue (*e.g.* Lafleur *et al.*, 2005; Ferrick and Price, 2009), a process understanding of shrubs in peatlands affected by wildfire has been limited to albedo measurements alone (Tsuyuzaki *et al.*, 2009). Compared to trees, the relatively rapid regrowth of the shrub canopy provides an interim vegetation cover for the peat surface that ameliorates the fire effects by providing shading and lowering apparent surface albedo. Unlike trees, the shrubs do not appear to contribute to snow ablation, meaning that recently burned peatlands are at an advantage in terms of snowmelt water inputs. While some previous work has examined the role of shrubs in mostly treeless peatlands (Admiral and Lafleur, 2007), this work is significant in that it is the first to contrast the ecohydrological processes of peatlands under the transition from a treed (unburned) to a shrub (burned) dominated system.

This work also contributes to the body of research on peatland ecology and post-fire vegetation succession, in that it provides a process understanding of the observed successional trends. The differential recolonization rates between hummocks and hollows noted previously in the literature can now be related to contrasts in the vadose zone hydrology, specifically the pore-water pressure distributions. Hummocks were found to have a greater and more variable pore-water pressure regime, which is due to a combination of distance from the water table and importantly the disequilibrium conditions. The harsher hydrological regime of hummocks after wildfire has implications in the perception of *Sphagnum fuscum* hummocks as "ecosystem engineers" (van Breeman, 1995) largely immune to stress and disturbance.

The water retention for peat presented here contributes relationships between simple peat properties such as bulk density, root content, and wood content in a framework that allows them to be related back to widely used water retention models (*i.e.* Van Genuchten, 1980). This work will facilitate the accurate modelling of peat moisture across a wide range of peat properties. Moreover, the observation of diminishing returns of peat moisture retention with increasing humification independent of changes in bulk density has important implications for the modelling of sustained smouldering combustion in deep peat, which largely depends on the ratio of moisture to peat mass (Benscoter *et al.*, 2011). This relationship may be key to the enhanced smouldering in peatlands subject to enhanced decomposition through long-term drainage or drying (Turetsky *et al.*, 2011).

6.3. Future work

Fortunately, work expanding on the findings presented here has already begun at the time of the submission of this thesis. James Sherwood has recently completed a M.Sc. thesis examining the ecohydrology of a forested fen that was drained and subsequently burned in the Chisholm wildfire of 2001. Dr. Nick Kettridge expanded on the simple ground heat flux work shown here by implementing a 3-D thermal model to burned and unburned hummocks and hollows. Steve Baisley is currently expanding on the transpiration work shown here by monitoring sapflow trends in black spruce trees affected by peatland drainage brought about by road construction. Beyond looking at drainage effects on transpiration, it would be worthwhile to examine trends in transpiration in peatlands of greater time since fire in order to test whether the shading-transpiration relationships hold over the entire fire cycle.

Under limited scenarios, boreal wildfires have been shown to be a net negative radiative forcing (Randerson *et al.,* 2006). The main contributor to the negative radiative forcing after wildfire in the boreal forest is albedo, particularly the transition from a low winter albedo in conifer forests to a high winter albedo in post-fire stands. In the case of the post-fire chronosequence studied by Randerson *et al.* (2006), the post-fire succession was from a black spruce to a deciduous stand for the first 40 years after fire, which shows a similar winter albedo to BC06. In forested peatlands, the delayed growth of conifers

(Wieder *et al.,* 2009) may infer a similar pattern of high albedo, but the effect may be muted by much higher CO_2 and CH_4 emissions during the wildfire itself. While this study provides albedo estimates before fire and within the first four years since fire, other factors such as enhanced black carbon emissions (Flanner *et al.*, 2007) may also counter the albedo effect. Post-fire albedo changes are not likely tied to fire severity in peatlands, but other positive radiative forcings such as black carbon and greenhouse gas emissions are tied to fire severity (Shetler *et al*., 2008). The work presented here on the water balance and peat characteristics of forested peatlands provides a foundation for the accurate estimation of emissions from wildfire, which is largely a function of moisture availability and retention (Benscoter *et al.*, 2011).

The relationships of basic peat properties to water retention models shown here could also be applied to larger databases of peat cores, such as that of Zoltai *et al*. (2000) and an upland soil properties database of Tarnocai and Lacelle (1996) that contains thin peat deposits. An improved understanding of the spatial (vertical and lateral) patterns of peat properties within the 1-2 m scale would assist in the prediction of peat combustion patterns during wildfire, as models such as Benscoter *et al.* (2011) require spatial data on peat density and water content. While maps of peatland extent exist at a coarse scale for Canada (Tarnocai *et al*., 2000), predictive models of peatland depth and vertical peat properties do not currently exist. In order to properly assess continent-scale emissions from wildfire inclusive of the deep-burning of peatlands, such future landscape-scale peatland models could be combined with the water retention relationships shown here.

While the work presented here focused on a peatland of average thickness (1.5 m) within the centre of a large bog island, there exists extensive thinner peat deposits whose properties and ecohydrological processes are poorly constrained in the literature. In ombrotrophic peatlands this is largely due to the more simplistic water balance possible at the apex of a domed bog (Ingram, 1982), such that most investigations are focused on the interior of the peatland. Bhatti *et al*., (2006), in a series of transects from the centre of a large fen to the upland margin, found a monotonic thinning of the peat with thinner peat deposits correlated with greater tree density. While models such as Ingram (1982) relate groundwater flow to the ratio of peat thickness to peatland width, other processes in thin peatlands are poorly understood. In particular, the hydrological response of thinner peat deposits to drought and drying (and the subsequent wildfire severity) may be significantly impacted by the water table dropping below the basal peat layers, thus exposing the unsaturated zone to very low conductivity clay deposits typically underlying peatlands. These drying conditions in thin peat deposits may also enhance the growth of trees due to increased aeration in the rooting zone. From the research presented here, the question emerges regarding whether peatland resilience to drought and wildfire is a linear function of peat thickness or if a critical thickness must be achieved for resilience to fire. Moreover, the question of peatland area and perimeter is also of future research interest, as hydraulic gradients can often move water from peatlands to uplands (Ferone and Devito, 2004). In conifers, seed throw decreases exponentially with distance from the tree, such that in smaller peatlands, more of the peatland is susceptible to being colonized by seedlings from uplands species such as jack pines, which show an ability to grow in sufficiently dry peatlands after wildfire (Pellerin and Lavoie, 2003), leading to enhanced transpiration fluxes. Investigation is worthwhile in determining if gross geometric characteristics of peatlands such as thickness and area to perimeter ratio can be used to estimate peatland resilience to wildfire.

6.4. References

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