Long Term Effects of Wildfire on Permafrost Stability and Carbon Cycling in Northern Peatlands

by

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Abstract

Changing fire dynamics and increasing global temperatures are causing changes to the fire regime and permafrost stability in the Arctic. Models have separately predicted the widespread thawing of permafrost and increasing magnitude and intensity of wildfires over the next century. However, while it is evident that wildfire and permafrost are both dominant controls on carbon dynamics in the boreal, less is known about the potential effects of wildfire to cause increased permafrost thaw and to affect soil carbon stocks. To assess the role of wildfire in future permafrost stability and carbon storage in permafrost peatlands of northwestern Canada, I addressed the following questions; 1) to what extent does fire accelerate permafrost thaw in discontinuous permafrost regions and 2) what is the magnitude of carbon released from depth as a result of fire-induced permafrost thaw on peat plateaus. This research was conducted at a variety of sites located within the Northwest Territories that burned 2,3,9,14,15,21,33,40 and 48 years ago, and six, nearby unburned sites. Field data was complemented with the use of remotely sensed data to determine the extent of fire-induced permafrost thaw. Soil respiration was measured on a subset of these sites. The results of these studies, summarized in this thesis, find that wildfire destabilizes the post-fire soil thermal regime of peat plateaus manifested by deeper active layers and widespread formation of taliks, persisting for up to 30 with the most pronounced effect being 10-20 following fire. This also appears to result in ~3 times the rate of recent thermokarst formation, making wildfire responsible for ~25% of permafrost thaw in the past 30 years. As a result of fire induced deepening of the active layer and increased soil temperatures, soil respiration at depth was stimulated, representing a nearly four times greater respiration of old compared to unburned sites. Surprisingly though, I find that this is not enough to offset the importance of surface labile carbon pools, to soil respiration, that are burned off during the fire event, despite the mobilization and release of deep, old, previously stored carbon. This thesis concludes by highlighting the importance of wildfire as a driver of permafrost stability and old carbon storage capabilities and emphasizes the importance of it in global permafrost and carbon cycling models.

Preface

This thesis is an original work of Carolyn Gibson. Chapter 1 provides an introduction to permafrost dynamics and carbon cycling in the Taiga plains and an overview of each thesis chapter. Chapter 2 has been formatted for submission to *Proceedings in the National Academy of Science*, and is currently *in preparation.* Chapter 3 has been formatted for *Journal of Geophysical Research: Biogeoscience* and is currently *in preparation*. Finally, Chapter 4 summarizes and synthesizes the finding from this work.

Co-authors made important contributions to each dissertation chapter. For chapter 2, L. Chasmer assisted with remote sensing sampling design and classification technical assistance, D. Thompson with the design and technical advice on measuring ground heat fluxes, and B. Quinton and M. Flannigan with concept formation and manuscript composition. D. Olefeldt was the supervisory author and assisted at all parts. For chapter 3, C. Estop Aragones assisted with sampling design and M. Flannigan and D. Olefeldt were supervisory authors.

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Table of Contents

Chapter 1: Introduction……………………………………………………………………1

1.1 Overview of permafrost peatlands…………………………………………………..…1
1.2 Permafrost and the development of peat complexes……………………………….....1
1.3 Wildfire and the Boreal Forest…………………………………………………………3
1.4 Interaction between wildfire and permafrost…………………………………………...4
1.5 Carbon cycling in the Boreal………………………………………………………….....5
1.6 Research Objectives……………………………………………………………………7
1.7 Literature Cited…………………………………………………………………………9

Chapter 2: Wildfire as a major driver of recent permafrost thaw in boreal peatlands……16

2.1 Abstract…………………………………………………………………………….16
2.2 Significance statement………………………………………………………………...16
2.3 Introduction……………………………………………………………………..…17
2.4 Results and discussion………………………………………………………………...18
2.4.1 Effects of wildfire on peat plateau soil thermal regimes……………………...18
2.4.2 Effects of wildfire on young thermokarst bog expansion…………………….21
2.5 Conclusion……………………………………………………………………………23
2.6 Materials and Methods……………………………………………………………….24
2.6.1 Site selection for soil thermal regime monitoring……………………………...24
2.6.2 Depth to frost table monitoring………………………………………………...24
2.6.3 Air and soil temperature monitoring…………………………………………...25
2.6.5 Image acquisition and preparation for supervised peatland classification……..25
2.6.6 Supervised peatland classification……………………………………………..26
2.6.7 Accuracy assessment of peatland classification……………………………….26
2.6.8 Estimating trajectories of young thermokarst development…………………...26
2.6.9 Spatial scaling of peat plateau loss due to thermokarst development…………26
2.7 Acknowledgements…………………………………………………………………...27
2.8 References……………………………………………………………………………27
2.9 SI Appendix…………………………………………………………………………...32
2.9.1 Site selection for soil thermal regime monitoring……………………………...32
2.9.2 Thaw depth monitoring………………………………………………………...32
2.9.3 Air and soil temperature monitoring…………………………………………...33
2.9.4 Vegetation Surveys………………………………………………………………33
2.9.5 Image acquisition and preparation for supervised peatland classification……..33
2.9.6 Supervised peatland classification……………………………………………33
2.9.7 Trajectories of young thermokarst bog development………………………….34
2.9.8 Scaling of regional peat plateau loss due to thermokarst expansion…………..35
2.9.10 Supplementary Literature Cited……………………………………...……55

Chapter 3: Effects of wildfire on soil respiration in permafrost peatlands of the Canadian Western Boreal Forest………………………………………………………………….……56

3.1 Abstract………………………………………………………………………………56
3.2 Introduction…………………………………………………………………………...56

3.3 Methods……………………………………………………………………….59
3.3.1 Study area……………………………………………………………….59
3.3.2 Field data collection……………………………………………………..59
3.3.3 Modeling and statistical analysis…………………………………………61
3.4 Results……………………………………………………………………...62
3.4.1 Soil temperature…………………………………………………………...62
3.4.2 Soil respiration…………………………………………...……………..63
3.4.3 Soil respiration models…………………………………………………..63
3.4.4 Deep peat contribution to CO_2 release……………………………..…64
3.5 Discussion………………………………………………………………………64
3.6 Conclusion…………………………………………………………………..…67
3.7 Acknowledgements…………………………………………………………..…68
3.8 References………………………………………………………………………69

Chapter 4: Conclusions………………………………………...……………………84
4.1 Literature Cited……………………………………………………………………86

Bibliography…………………………………………………………………….........87

<u>List of Tables</u>

Table S2.1: Summary statistics from chronosequence sites. Date and size of fire derived from Large Fire Data Base (NRCAN 2016). Number of active layer measurements include only locations used in analysis, ie. hollows excluded…………………………………………….....42

Table S2.2. Dating of transitions between plateau peat and young thermokarst bog peat in cores collected from current young and mature thermokarst bog sites. All dating is done ~1-2 cm below the transition, which means that true transition dates happened more recently than indicated by the dating techniques. Note that these cores were not collected at the transition between young and mature thermokarst bogs, but rather centrally within young and mature thermokarst bogs, respectively, and thus in order to assess the persistence of young thermokarst bogs in the landscape, we are interested in seeing the maximum age of transitions within current young thermokarst bogs and the minimum age of transition in current mature thermokarst bogs………………………………………………………………………………………... 43

Table S3.3. Summary statistics from peatland classification showing sites with site coordinates, number of 250 x 250m sections classified and percent coverage of the three peatlands classes ± one standard deviation…………………………………………………………...………………44

Table S2.4. Accuracy assessment for predictions of absence or presence of taliks within a burned peat plateau site using September thaw depths. Data is from one unburned site, Lutose Unburned (See table S2.1), where actual absence or presence of taliks was determined by seasonally monitoring of thaw depths……………………………………………………….……………45

Table S2.5. Accuracy assessment for predictions of absence or presence of taliks within a burned peat plateau site using September thaw depths. Data is from one burned site, Lutose 2012 (See table S2.1), where actual absence or presence of taliks was determined by frequent monitoring of thaw depths……………………………………………………………………………….....45

Table 3.1: Modeled equations for burned and unburned sites, arranged in order of decreasing model fit based on Akaike information criterion………………………………………………80

Table 3.2: Coefficients (±1SE) of variables in the best model based on an AIC evaluation……80

List of Figure

Fig. 2.1. Wildfire in western Canadian boreal peatlands. A) Outline of the study region, defined as the discontinuous permafrost zone (21) within the Taiga Plains ecozone (22). Shading within the study region shows the coverage of peat plateaus as indicated by histel soils (23), while historical burned areas from the last 30 years are shown as hashed areas (24). Yellow stars indicate locations where field data characterizing soil thermal regimes were collected, with one unburned and at least one burned site monitored at each star location. Blue stars indicate locations where satellite image classification of recent thermokarst bog expansion was carried within extensive peatlands partially affected by wildfire. B) Example of a peatland characteristic for the study region, with three open non-permafrost thermokarst bogs surrounded by treed permafrost-affected peat plateaus. Distinct young thermokarst bog stage is visible as green areas along some peat plateau edges, a result of recent permafrost thaw. Photo copyright Mason Stothart. C) Example of a characteristic crown fire burning a peat plateau, near Fort Simpson, NWT, in 2013. Photo copy right Franco Alo………………………………………………..…………….36

Fig. 2.2. Simplified illustration of soil thermal states for peat profiles along transects from thermokarst bogs to peat plateaus within and outside historical burned areas………………………………………37

Fig. 2.3. Vegetation recovery at select peat plateau sites used in this study to assess long term effects of fire on soil thermal regime……………………………………………………………………………...38

Fig. 2.4. Trajectories of soil thermal regime and vegetation recovery on peat plateaus following wildfire. A) Trajectories of depth to frost table at point without taliks, and proportion of points where thaw depth thaw monitored that had continuously thawed soil layers, i.e. taliks. No depths to frost table estimates were possible for sites burned 7 and 16 years ago due to 100% talik coverage. B) Trajectories rate of that, as indicated by soils at 40cm rising above 0.5°C, and maximum seasonal soil temperature attained at 40 cm depth (points at 7 and 49 years post fire not available). C) Trajectories of vegetation recovery, exemplified by the percent ground cover of lichens, and median height of black spruce trees. In each panel, the average of 6 unburned sites is shown on the left with grey background, where error bars indicate ±1 standard deviation for percent lichen and interquartile range for black spruce trees…………39

Fig 2.5. Classification of peat plateau, young thermokarst bog, and mature thermokarst bog extents in large peatlands using high resolution satellite imagery (World View 2, 50 cm resolution). A) and C) Supervised classification was carried out for 250 x 250 m sections of unburned burned peatlands respectively that were visually assessed to only contain peat plateau, young thermokarst bog, and mature thermokarst bogs. Sections containing channel fens, upland forests, ponds, clouds, or shadows were excluded from analysis and are shown with grey shading. B) and D) Example of a 250 × 250 m section selected for classification in the unburned and burned peatlands respectively……………………………40

Fig 2.6. Fig 6. Effect of wildfire on permafrost thaw through development of young thermokarst bogs in boreal western Canada. A) Current-day coverage of young thermokarst bogs within and outside 20- to 30-year-old fire scars at four large peatlands. Box plots indicate median, interquartile range, and minimum and maximum young thermokarst bog coverage among classified 250 × 250 m sections, and number of classified sections is indicated (n). In each pair, burned sections had significantly greater was significantly different from its unburned pair (*$p<0.01$,* pairwise t-test). Sites are ordered left to right by decreasing mean annual air temperature. B) Temporal development (± 95% CI) of young thermokarst bogs currently present inside and outside 30-year-old fire scars. Young thermokarst bogs persist in peatlands for 100 ± 40 years before succession into mature thermokarst bogs (15), suggesting that a significant proportion of the currently present young thermokarst bogs formed prior to the historical wildfire. Rate of development of young thermokarst bog, expressed as a % of total peatland area (± 95% CI), shown above each trajectory…………………………………………………………………….41

Figure S2.1. Example of a peat plateau site (Kakisa 2014) where soil thermal regime was monitored. Each peat plateau site had a 5 m spaced grid with 100 point locations marked with flags where depth to frost table was measured……………………………………………………..46

Fig. S2.2. Example of transects from peat plateau, through young thermokarst bog, to mature thermokarst bogs. (a) Lutose 2012, b) Samba deh 2013, c) Fort Simpson unburned, d) Lutose unburned. Notice the shift in dominant vegetation between the young and mature thermokarst bogs……………………………………………………………………………………………47

Fig S2.3. Determination of talik presence, and estimation of thaw depth at point locations where taliks were absent. A) Example of seasonal development of the thawed soil layer at 6 representative point locations at the Lutose 2012 site. Soil probe was 150 cm long, hence measures of 150 cm indicate that no frost table was reached. We classified point locations as having taliks if there was a >90 cm increase in thaw depth between two monitoring occasions, as exemplified by three point locations. Only a subset of sites had thaw depths monitored as frequently as Lutose 2012, and thus for other sites we did not have direct observations of talik frequency. B) Histogram of September point location depth to frost table measurements (n = 100) from the Fort Simpson 1995 site overlaid by a gaussian kernel density estimation (Wessa 2015). All point locations with September thaw depths >100 cm were assumed to have taliks, an assumption that was tested against observed talik presence at a few sites (See table S2 and S3). We used the maximum density to estimate representative active layer depth of point locations where taliks were absent, as using either the mean or median thaw depths would be strongly influenced by talik frequency………………………………………………………………….48

Fig S2.4. September active layer depth with mean annual air temperature in the unburned sites in the burned and unburned sites. Climate was not found to be the main influence on active layer depth, supporting of conclusion that wildfire in the main driver of difference in active layer depth…………………………………………………………………………………………...49

Fig S2.5. Differences in the seasonal development of the thawed soil layer between unburned and historically burned sites. A) Thaw depths are shown ±1 standard deviation of n=3 (2-5 years), n=4 (7-22 years), and n=3 (32-55 years). Point locations where taliks later developed were included for June and July dates. B) Absolute difference in depth to frost table between burned and paired unburned sites in early June. Sites 10, 21, 32 post fire not included due to inaccessibility at this time of year………………………………………………………………..50

Fig S2.6. Example section of World View 2 satellite image. Supervised classification was carried out for 250 x 250 m sections of peatlands that were visually assessed to only contain peat plateau, young thermokarst bog, and mature thermokarst bogs. Sections containing channel fens, upland forests, ponds, clouds, or shadows were excluded from analysis and are shown with grey shading…………………………………………………………………………………………51

Fig S2.7. Field validation of extents of young thermokarst bog. World View 2 Satellite Image (75 × 75 m) shown right with corresponding classification on the left. Overlain points taken with dGPS unit, up to 10 cm accuracy……………………………………………………………...52

Fig. S2.8. Residual difference between edges of supervised classification and field measurements of edge. Overestimates occurred when classification is outside the field delineated young thermokarst bog area, underestimates occurred when the classification was within the field delineated young thermokarst bog area…………………………………………………………53

Figure S2.9. Estimate of the trajectories of young thermokarst bog development. A) Trajectories of recent thermokarst bog development based on assumptions 1-4. B) 95% confidence intervals for each site, from which mean and standard deviation of the mean were used for final trajectory estimate……………………………………………………………………………………..54

Figure 3.1: Peatland complexes and field sited in the Taiga Plains ecozone (inset) in western boreal Canada. Peatland plateaus within the Taiga Plains are a mosaic of treed permafrost peatlands and open, permafrost free thermokarst bogs. Study sites included two burned permafrost peat plateaus that burned during the summer of 2000 and 2007, and an unburn site…………………………………………………………………………………......75

Figure S3.1: Soil moisture using a handheld soil moisture sensor and soil respiration. No relationship was found between soil respiration and moisture therefore soil moisture was not included in our models of soil temperature as a controlling factor………………………………76

Figure S3.2: Soil temperature, in two hour intervals, from hobo temperature loggers from shallow, 10cm (upper lines), and deep, 40cm (lower lines), during growing season. Temperature is averaged from the two loggers located within each site………………………………………77

Figure 3.2: Soil temperature during growing season (JD 150-250). A) Soil temperatures at 10cm and were significantly warmer in the burned than in the unburned ($t= 2.705, p<0.05$). B) Soil temperatures at 40cm were also significantly higher in the burned than in the unburned ($t=5.414,p<0.05$) during the 2016 growing season……………………………………………..78

Figure 3.3: Raw soil respiration during the growing season (JD 148-300). A) Soil respiration during the season with a second order polynomic trend line. Unburned sites displayed higher respiration during the early summer while burned sites showed increased respiration late summer. B) Exponential relationship between soil temperature and soil respiration……………79

Figure 3.4: Modeled and measured soil respiration during the growing season (JD 140-300) in the burned (a) and unburned (b) sites. Model accurately predicted fluxes in the burned (R^2=0.79) and the unburned (R^2=0.71). Model performed most poorly at low temperatures and low fluxes, likely due to low n size at these conditions……………………………………………………81

Figure 3.5: Fractional contribution of deep respiration to total soil respiration from the burned and unburned sites. Burned sites displayed increased respiration from depth later into the summer season whereas the unburned site remained relatively consistent, with only a minor increase……………………………………………………………………………………..82

Figure 3.6: A) Cumulative carbon loss (gC m^2) from shallow soils (10cm) during the growing season (JD 150-258) with upper and lower bounds based on a Q_{10} of 2.56 ± 0.31. B) Cumulative carbon loss from deep soils (40cm) during the same period……………………………………83

Chapter 1: Introduction

1.1 Overview of the Importance of Permafrost Peatlands

Peatlands only cover 3-5% of the world's land area but store ~30% of terrestrial soil carbon [*Gorham 1991*]. Within northwestern Canada, permafrost is found almost exclusively in peatlands in the southern limit of the discontinuous permafrost zone. Approximately 28% of peatlands in continental western Canada are underlain by permafrost and store ~13 Pg of soil carbon. Given that peatlands 1) currently serve as a large terrestrial carbon sink and 2) are located primarily at northern latitudes that are undergoing rapid and accelerated warming associated with climate change, understanding their response to disturbances is vital for continued benefits derived from these ecosystems.

Peatlands within the discontinuous permafrost zone not only serve as a large terrestrial carbon sink, they also supply cultural resources such as traditional medicines, travel corridors, serve as headwaters for numerous river systems and help maintain stable wildlife populations [*Callahan et al., 2005, Walsh et al., 2005*]. When these systems undergo thawing events that can lead to either rapid thaw settlement or thermokarst development it can result in the ground instability which has important implications for land processes and hazards such as surface hydrology, groundwater regimes, and stability. These changes have further reaching consequences for socio-economics of northern communities, infrastructure and development projects [*Savo et al.*, 2017].

It is predicted that by the mid-21st century, the area of permafrost in the northern hemisphere will decline by 20-35 per cent [*IPCC 2016*]. Permafrost peatlands near the southern extent of the permafrost zone are most likely to be impacted first, thereby triggering a cascade of potentially detrimental effects for the global carbon cycle and ecosystem services derived from permafrost peatlands. Therefore it is important to focus research efforts on understanding mechanisms that may trigger or accelerate permafrost thaw in order to make informed management and development decisions in the future.

1.2 Permafrost in Western Canada and Development of Peatland Complex

Across western Canada 21%, or over 365,000km^2, of the land is covered by peatlands [*Vitt et al.*, 2000]. In this region, peatlands are a heterogeneous mosaic of peatland types including ombrotrophic bogs, minerotrophic fens, and permafrost features (peat plateaus). Permafrost underlays approximately 28% of these peatlands [*Tarnocai et al.*, 2004] and is located within the discontinuous permafrost zone where it is restricted mostly to ombrotrophic peatlands. In these areas peat plateaus arise because the rate of production of organic matter exceeds losses from decomposition, disturbance, and leaching [*Vitt et al.*, 1995] and as the result of being ice rich, these features are elevated 0.5-1.5m above surrounding wetlands [*Quinton et al.*, 2011]. These areas are distinguishable by the presence of black spruce (*Picea mariana)* that is able to establish due to the creation of an aerobic active layer in the upper part of the ground due to summer seasonal thaw.

Many studies have shown that on century time scales both the aggradation and degradation of plateaus can occur within the same peatland [e.g. *Zoltai*, 1972; *Payette et al.*, 1976]. Permafrost is maintained in peatlands due to a hydrological gradient that exists between forested plateaus and surrounding wetlands (bogs and fen) such that adjacent wetlands receive relatively large inputs of runoff from plateaus [*Quinton et al.*, 2009]. The ground surface on peat plateaus is dominated by bryophytes and the insulating properties provided by dry surface peat, mosses and lichens are vital for the persistence of permafrost in peatlands. In the winter, given thin enough snow layers, heat is lost through the frozen peat and in the summer, permafrost is insulated from increasing air temperatures due to the presence of *Sphgnaum fuscum* hummocks and *Cladonia rangiferina* [*Zoltai*, 1993], which reduces the heat flux downward towards the permafrost [*Seppala 1988*].

However, degradation of plateaus can be triggered by both natural and catastrophic events. As ice accumulates, a plateau will continue to increase in height to a point where the peat can no longer cover the surface [*Seppala 1988*]. Once this occurs, surface cracks begin to form, creating deep fissures allowing warm air and rain to reach the frozen core and initiate thawing [*Zoltai*, 1972]. As thawing progresses, the land subsides, increasing the saturation of the surface peat as it begins to fall into or below the rooting zone. Black spruce, the dominant tree species on peat plateaus, cannot tolerate these waterlogged, low-oxygen conditions [*Islam and Macdonald*, 2004] in the rooting zone and begin to die. Loss of black spruce is accompanied by *Sphagnum fuscum*, characteristic of the dry peat plateaus, being gradually replaced by *Sphagnum* *angustifolium* then by *Sphagnum riparium* as subsistence continues and the water table approaches the surface of the peat [*Gignac et al.*, 1991; *Zoltai*, 1993]. This thawing process, under some circumstances, can be initiated by wildfire but historically it was believed that fire did not cause widespread permafrost thaw in peatlands [*Thie*, 1974] and that the rate of aggregation and degradation of permafrost were in balance [*Zoltai*, 1993]. This process is referred to as thermokarst formation and while it is natural, it is predicted that our changing climate may accelerate rates of thermokarst formation. Over the latter half of the 20th century, the area of permafrost peatlands decreased by 22% on average [*Beilman and Robinson*, 2003]. This conversion of permafrost plateaus to permafrost-free bog and fens will have drastic implications for vegetation succession, carbon storage, and ecosystem services derived from these areas [*Turetsky et al.*, 2007].

Permafrost thaw can also occur in the regions in the form of top-down thaw. The active layer, the layer of soil between the permafrost and the surface that thaws each summer due to increased air temperatures, can become progressively deeper each summer. In some circumstances, seasonal thaw becomes so great that due to the quick onset of winter not all of the energy stored in the thawed soil is able to be evacuated. This can result in a perennially unfrozen peat layer between the permafrost table and the frozen upper layers called a talik. The soils within a talik remain above 0°C thereby acting as a year-round heat source for further active layer deepening [*Viereck et al.*, 2008].

1.3 Wildfire and the Boreal Forest

Although not historically common, peat stratigraphy and macrofossil analysis has shown evidence of some wildfire-induced permafrost thaw; a transition from silvic peat (decomposed woody material) to *Sphagnum* mosses is preceded by a char layer. Wildfire is a major stand-renewing agent for much of the circumboreal forest zone and is an integral part in determining patterns of vegetation and forest structure within the boreal [*Rowe and Scotter*, 1973; *Stocks et al.*, 2002]. Wildfire within peatlands represents a large source of carbon to the atmosphere, burning over 1500km^2 and releasing 6GtC annually in western Canada alone [*Turetsky et al.*, 2002] on a 100-120 year fire return interval [*Turetsky et al.*, 2004]. Peatlands have been considered relatively resistant to wildfire due to their moist soil conditions [*Kuhry*, 1994] but recent studies have shown that soil combustion in peatlands during wildfire can be substantial both in extent [*Turetsky et al.*, 2004] and severity [*Turetsky and Wieder*, 2001; *Turetsky*, 2002; *Benscoter and Wieder*, 2003; *Turetsky et al.*, 2011]. Wildfire affects forested peat plateaus in similar frequencies to uplands forests (100-120 fire return interval; *Turetsky et al. 2004*) and in the last 50 years more than 25% of peat plateaus within the Taiga Plains have burned. Changes to the fire regime are expected to occur in the future under a predicted changing climate that could drastically alter fire dynamics on the landscape and potentially increase these numbers. Climate change has altered the boreal fire regime and these changes are predicted to become more pronounced in the future [*Kasischke and Turetsky*, 2006; *Flannigan et al.*, 2009]. Across North America, the total annual burn area has increased due to higher mean annual temperatures and drier conditions [*Kasischke and Turetsky*, 2006]. Fire occurrence is currently predicted to increase by 25% by the year 2030 and 75-150% by the end of the century [*Wotton et al.*, 2010].

1.4 Interaction between Wildfire and Permafrost

Wildfire is an important driver of ecosystem structure. Within the discontinuous permafrost zone, by combusting aboveground biomass and some of the uppermost organic material, wildfire can dramatically alter the land surface energy balance by exposing soils and reducing the insulation effect of organic layers. This causes modifications to the ground thermal and hydrological conditions that can affect permafrost stability [*Yoshikawa et al.*, 2002]. Several studies have begun to explore the effects of wildfire on permafrost stability and have noted impacts from 1 to more than 40 years after fire [e.g. *Burn*, 1998; *Yoshikawa et al.*, 2002; *Viereck et al.*, 2008; *Brown et al.*, 2015; *Zhang et al.*, 2015]. Studies show that fire can increase soil temperatures for several decades and the active layer usually deepens [*Mackay*, 1995; *Burn*, 1998; *Viereck et al.*, 2008; *Rocha et al.*, 2012; *Nossov et al.*, 2013]. Some observations show permafrost table stabilization or recovery 5–25 years after a fire [*Mackay*, 1995; *Yoshikawa et al.*, 2002; *Smith et al.*, 2015]; others show steady deepening of the permafrost table three to six decades after a fire or removal of vegetation and surface organic matter [*Burn*, 1998; *Viereck et al.*, 2008] and even complete degradation of permafrost after a fire [*Viereck et al.*, 2008]. Several modeling studies have shown that vegetation recovery and the accumulations of organic material are important for stabilization and recovery of the permafrost table [e.g. *Jafarov and Romanovsky*, 2013; *Smith et al.*, 2015].

There is sufficient evidence to suggest wildfire plays a role in permafrost conditions, however, presently the effects of fire are not considered in traditional permafrost maps [*eg. Brown et al. 1997, Heginbottom et al. 1995*]. Furthermore, global- and national-scale permafrost modelling studies do not consider the effects of fire, often due to their coarse spatial resolution [*Zhang*, 2005; *Koven et al.*, 2015]. Although fine scale studies have recognized the importance of fire on soil thermal conditions and show significant increases in near-surface soil temperatures and active layer thickening, they have not made the connection between potential active layer thickening and thermokarst development. These two phenomena have only been explored in isolation. Furthermore, most of the research conducted has been done in Alaska where peat depth is often <1m thick and on north-facing hillsides rather than in larger lowland peatland complexes [e.g. *Viereck et al.*, 2008; *Brown et al.*, 2015], much less than that found in the western boreal peatlands of Canada. These studies suggest that the response of permafrost to wildfire is mediated by fire severity (% organic layer consumption). For example, Nossov et al. (2013) argue that when less than 4cm of organic material remains post fire it is likely to lead to complete permafrost loss. However, the vulnerability of permafrost to wildfire is likely to differ across the discontinuous permafrost zone depending on numerous site factors that influence soil heat flux. The current literature neglects the wetland rich areas of northwestern Canada where permafrost is found under slightly raised (0.5-1.0m), tree covered plateaus that overlay frozen organic soils up to and >10m in depth [*Wright et al.*, 2009]. Understanding permafrost response to wildfire in this region is important as it contains over 20% of the circumpolar soil organic carbon stock and approximately a quarter of these permafrost soils have burned in the last 30 years.

1.5 Carbon Cycling in the Boreal

Northern peatland ecosystems accumulate carbon because annual net primary productivity (NPP) of peatland vegetation generally exceeds lateral transport and the annual decomposition of litter and peat. Over millennia NPP has been greater than decomposition and hence, northern peatlands have been a persistent sink of CO_2, averaging 0.02 to 0.03 kg CO_2-C m^{-2} y^{-1} [*Gorham*, 1991; *Tolonen and Turunen*, 1996] . One of the most significant knowledge gaps related to the northern permafrost regions is the impact thawing of frozen soils will have on the global carbon cycle. Some researchers believe the effect may be catastrophic, while some are skeptical about its significance. Understanding the mechanisms that control aerobic and anaerobic respiration pre-thaw and post-thaw at different spatial and temporal scales is particularly important for predicting the effects of changing disturbance regimes and a changing climate on northern carbon cycling.

Following permafrost thaw, herein referring to active layer deepening for purposes of this thesis, large quantities of labile carbon, which were previously resistant to decomposition due to subzero temperatures, become available for decomposition, resulting in the transfer of old carbon from the terrestrial environment to the atmosphere. The rate and timing of this, however, remains a key uncertainty [*Schuur et al.*, 2015]. Variations in post thaw trajectories (active layer deepening versus thermokarst creation) can drive differences in CO_2 production rates due to differences in oxygen conditions [*Schuur and Bockheim*, 2008]. In contrast to thermokarst formation that results in anaerobic conditions, active layer deepening typically exposes thawed soil organic carbon stores to aerobic conditions [*Lee et al.*, 2012]. These aerobic conditions are suspected to be particularly important in controlling the rate of CO_2 transfer to the atmosphere because decomposition that uses oxygen as the terminal electron acceptor is 5 to 10 time faster than anaerobic decomposition [*Bridgham et al.*, 1998]. Furthermore, rates of soil respiration can further be expected to increase as soil temperatures post fire are greater due to changes in the surface energy budget that results from the removal of the vegetation layer and exposure of a predominantly black charred surface [*Randerson et al.*, 2006]. This increase in soil temperature is particularly important as in peat, carbon dioxide production rates increase by a factor of 2-3 for every 10ºC temperature increase (Q_{10}) [*Moore and Dalva*, 1993; *Yavitt et al.*, 1997; *McKenzie et al.*, 1998].

Decomposition of thawed permafrost soils, whether due to thermokarst formation or active layer deepening, will result in the decomposition of old carbon that had been previously protected from decomposition. Previous work examining the release of old carbon following permafrost thaw have shown that following thermokarst development up to 15% of ecosystem respiration is derived from the decomposition of old carbon [*Schuur and Bockheim*, 2008]. However, most work examining the decomposition of previously frozen carbon stores has examined rates of decomposition following thermokarst development. Much less studied is the potential for old carbon losses associated with deepening active layers, particularly following wildfire. Up to a 200% increase in active layer depth has been observed following fire [*O'Neill et al.*, 2002], representing a vast pool of organic matter that is now available for microbial decomposition and could represent a large pulse of carbon to the atmosphere from the terrestrial environment. The magnitude and timing of this decomposition is not well understood. Furthermore, previous studies that have attempted to understand this have focused primarily on black spruce upland forests, which tend to have shallower organic layers, and examined net ecosystem respiration. While this provides a good approximation for trends in CO_2 production following a wildfire, it provides little insight into the source of carbon. It is unclear if observed increases in ecosystem respiration are due to increased soil temperatures in the upper peat surfaces or due to loss of old carbon at depth.

1.6 Research Objectives

The overall goal of my research is to document the temporal and spatial time frame over which wildfire affects permafrost stability and carbon cycling in the discontinuous permafrost zone. Despite previous studies showing changes to the surface energy balance post fire [*Randerson et al.*, 2006; *Thompson et al.*, 2015] and deepening of active layers [*Viereck et al.*, 2008; *Brown et al.*, 2015], no studies have explored the link between these factors and widespread permafrost thaw. Using a chronosequence approach (space for time substitution), I aim to understand the spatial extent of wildfire induced permafrost thaw and the fate of soil carbon post-fire by asking the following questions:

1. Over what temporal scale does wildfire affect active layer depths?
2. To what extent does fire accelerate thermokarst development in discontinuous peatland-rich regions?
3. How is soil respiration affected over the years to decades following wildfire in peat plateaus?
4. Relative to unburned peat plateaus, what is the magnitude of post-fire CO_2 respiration and what proportion of this respiration results from the decomposition of deeper, recently thawed C stores?

1.5.1 Wildfire and Permafrost Stability

Preliminary investigations have explored the impacts of wildfire on active layer deepening in Alaska where organic layer depths are often <1m. In chapter 2, I determine if and to what extent fire induced active layer deepening leads to thermokarst formation. Active layer depths were measured across a series of sites that burned from 1967-2014. These sites were also fitted with several temperature loggers to monitor seasonal variation in soil temperatures with time since fire. This allowed me to assess the temporal scale over which wildfire impacts active layer depths in peat rich areas such as the Taiga Plains. I was then able to use high resolution satellite imagery with supervised classification to determine if the area of thermokarst varies between adjacent burn and unburn areas. As a result of wildfire on the surface energy balance, and therefore the soil thermal regime, I expect that 1) active layer depths will be deeper during the first decade, followed by a decade of widespread talik formation, 2) soil temperatures at depth will be greater due to a greater ground heat flux, and 3) as a consequence of (1) and (2) thermokarst formation will be accelerated within a burned area compared to an adjacent unburned area.

1.5.2 Post-Fire Carbon Cycling

Changes in carbon emissions post thaw has largely been associated with increased methane respiration due to the creation of anaerobic conditions within thermokarst features and the respiration of old carbon [*Schuur and Bockheim*, 2008; *Schuur et al.*, 2015]. Chapter three deals with changes in carbon cycling caused by the deepening of the aerobic zone (active layer) due to wildfire. Using CO_2 fluxes from dark static chambers on an unburn site and sites that burned in 2000 and 2007 allowed me to document the magnitude of soil respiration both temporally and in comparison to an unburned site. Because wildfire increases the availability of soil organic carbon and increases the temperatures of the soils, I expect that 1) soil respiration in burned sites will be greater compared to unburned sites, and 2) burned sites will have proportionally more respiration that is derived from deeper soils due to higher soil temperatures at depth.

Chapter four will summarize the most important findings, analyze the strengths and weaknesses of this study, and provide direction for future research in relationship to wildfire and permafrost stability. Integrating the information contained with chapters two and three, I provide a discussion on possible future permafrost-fire-carbon scenarios in a changing climate.

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Chapter 2: Wildfire as a major driver of recent permafrost thaw in boreal peatlands

2.1 Abstract

Permafrost vulnerability to climate change may be underestimated unless effects of wildfire are considered. In this study we assessed decadal impacts of wildfire on soil thermal regime and abrupt permafrost thaw in western Canadian boreal peatlands. Using 16 sites that vary in time since fire, we found a near doubling in the depth to frost table in the year following fire while continuously thawed soil layers above the permafrost, taliks, subsequently became ubiquitous a decade after fire. The soil thermal regime slowly recovered with the reestablishment of black spruce and lichens, and effects of wildfire could not be distinguished at sites that burned >30 years ago. Given the perturbation of the soil thermal regime, we hypothesized accelerated abrupt permafrost thaw along permafrost edges, i.e. thermokarst bog expansion, during the first three decades following wildfire. To test this we quantified extents of developmentally young thermokarst bogs in four large peatlands partially affected by fire in the 1980s or 1990s. Young thermokarst bogs were nearly twice as abundant within fire scars, indicating a tripling in the rate of expansion after fire. Our results suggest over the last 30 years, wildfire is directly responsible for 2,200 ± 1,500 km^2 (95%CI) of thermokarst bog expansion, and thus responsible for ~25% of all recent permafrost thaw in this region. Permafrost thaw has major implications for traditional land-use, regional hydrology and water quality, and soil greenhouse gas emissions, and our study stresses the recent and future role of wildfire for permafrost stability in circumpolar boreal regions.

2.2 Significance statement

A quarter of the global land surface has permanently frozen ground, permafrost. Ongoing climate warming is causing rapid permafrost thaw, which has important effects on infrastructure integrity, traditional land-use, soil carbon storage and more. In this study we show that permafrost thaw in boreal regions is significantly accelerated in the first three decades following a wildfire. We estimate that wildfire is directly responsible for ~25% of all permafrost thaw over the last 30 years in a region in western Canada. Hence, projections of future permafrost thaw are likely underestimates unless they incorporate the effects of wildfire.

2.3 Introduction

Permafrost thaw has impacts at local to global scales, affecting infrastructure integrity (1), traditional land-use (2, 3), wildlife habitat (4), hydrology, and water quality (5, 6), and soil greenhouse gas emissions (7). Projections of 21st century permafrost stability indicate particularly rapid thaw in the discontinuous permafrost zone (8, 9), but projections may yet be underestimates as they do not explicitly consider potential destabilization of permafrost due to the role of disturbances, the largest of which in the boreal forest is wildfire. Wildfire affects 2.5 Mha of the boreal biome on average each year (10). Peat plateaus are ubiquitous in the boreal forest, and often burn preferentially due to the black spruce trees. Peat stratigraphy in boreal peatlands show that historical wildfires can cause permafrost thaw (11), but does not give information on the duration of wildfire's effects on soil thermal regimes or the spatial extents of permafrost thaw arising due to wildfire. Understanding the role of wildfire as a driver of permafrost thaw is critical given the recent, and projected, increases in annual burned areas in boreal regions (12).

This study focuses on the discontinuous permafrost zone within the Taiga Plains ecozone in western Canada, a 400,000 km^2 region that is representative of circumpolar boreal regions with extensive peatland cover (Fig. 2.1a). Here, peat accumulation initiated following deglaciation ~9000 years ago (13), and current peat depths are between 2 and 6 m. As such, this region has some of the highest soil organic carbon concentrations globally (14). Permafrost started aggrading after the Holocene thermal maximum ~5,000 years ago, and became more widespread following a climate cooling ~1,200 years ago (15). Permafrost is largely restricted to peatlands in the study region, due to the insulating properties of peat and the relatively warm mean annual temperatures between -1°C and -6°C. A relatively warm climate, and shallow permafrost basal depths less than 20 m (16), suggest that even minor perturbations to the surface energy balance can cause complete and rapid thaw.

Peatlands in the study region are a fine-scale mosaic dominated by permafrost-affected peat plateaus and non-permafrost thermokarst bogs (Fig 2.1b). Peat plateaus are elevated 1 to 3 m above their surrounding peatlands due to excess ground ice, causing generally dry surface conditions and vegetation dominated by black spruce (*Picea mariana*), Labrador tea shrubs (*Rhododendron groenlandicum*), and lichens (*Cladina spp.*) (Fig. 2.2). Black spruce forests burn preferentially in boreal western Canada (17), and ~ 25% of peat plateaus in the study region have burned in the last 30 years (Fig. 2.1a and Fig S2.1). Thermokarst bogs form through abrupt permafrost thaw, often developing from small depressions on peat plateaus which then expand radially and coalesce. Abrupt permafrost thaw cause peat surface collapse, which initially leads to a young thermokarst bog stage with saturated soils and vegetation dominated by *Sphagnum riparium* and sedges (*Carex aquatilis*). Vegetation composition shifts to *Sphagnum fuscum* and ericaceous shrubs as new peat accumulates and the surface becomes drier (Fig 2.1b and Fig 2.2, *SI Appendix,* Fig. S2.1). Repeat aerial photography and satellite image acquisition have shown that the rate of thermokarst bog expansion has accelerated in the last few decades due to warming (18–20), but no analysis has been done to compare rates of thermokarst bog expansion within and outside historical fire scars.

The objective of this study was to assess impacts of wildfire on permafrost stability in peat plateaus. We hypothesize that wildfire causes decadal perturbations to the surface energy balance, resulting in higher soil temperatures and a thicker active layer. We further hypothesize that the presence of a seasonally unfrozen layer during the winter further perturbs the soil thermal regime which renders peat plateau edges increasingly vulnerable to abrupt permafrost thaw through thermokarst bog expansion (Fig. 2.2). In order to test our hypotheses, we monitored soil thermal regimes at 16 peat plateau sites, ten of which had burned at various times in the last 50 years (Fig. 2.1a and Fig 2.3, *Supplementary Appendix*, Table S2.1), and we assessed extents of recently developed young thermokarst bogs in four large peatlands partially affected by fire 20 to 30 years ago (*Supplementary Appendix*, Table S2.2). By combining field measurements and remote sensing approaches, we are able to estimate the area and relative importance of permafrost thaw in the study region directly occurring as a consequence of recent wildfires.

2.4 Results and Discussion

2.4.1 Effects of wildfire on peat plateau soil thermal regimes

Soil thermal regimes at each of the 16 peat plateau sites (Fig 2.1a, *Supplementary Appendix,* Table S2.1) were characterized by monitoring air and soil temperatures at 10 and 40 cm depths for a full year and by repeat measurements of depth to frost table up to 7 times per site between May and late September of thawed soil layer depths at 100 point locations in a 5 m spaced grid (see methods). By tracking thaw depth development throughout the season, it was apparent from sudden large increases in thaw depths that many point locations had continuously thawed soil layers between the permafrost and the seasonally frozen layer, i.e. taliks (Fig. 2.2, *Supplementary Appendix*, Fig S2.3a). Using the within-site distribution of depth to frost table measurements in late September, we derived two indices for characterizing site soil thermal regimes; the proportion of point locations underlain by taliks, and the maximum depth to frost table for point locations not underlain by taliks (*Supplementary Appendix*, Fig S2.3b). All sites except one had peat depths > 1.9m, uniform pre-fire tree density, and mean annual air temperatures among the six unburned sites were between 0.8 to -3.1°C (*Supplementary Appendix*, Fig S2.4 and Table S2.1), hence differences in soil thermal regimes are considered likely to be primarily due to differences in fire histories.

We observed nearly 50% deeper depths to the frost table in recently burned sites, with ~50 cm in unburned sites compared to ~85 cm in sites burned in the last 5 years (Fig. 2.4a). We associate this phenomenon predominantly to a faster thaw rate throughout the summer (*Supplementary Appendix*, Fig S2.5a and Fig S2.5b), as opposed to difference in timing of spring thaw initiation or late season thawing dynamics, though these factors likely play a small role as well. This was supported by seasonal thaw at 40 cm depth occurring two weeks earlier in recently burned sites compared to unburned sites (Fig 2.4b). More rapid and deeper development of the seasonally thawed layer likely stems primarily from a perturbed surface energy balance with increased incident shortwave radiation following the loss of the black spruce canopy and reduced albedo due to the replacement of white lichen with black char (Fig 2.3 and Fig 2.4c). Interestingly, in southern boreal permafrost free bogs, wildfire causes reduced soil moisture that insulates the soil profile thereby counteracting the effects of albedo and increased incident shortwave radiation (25). However, permafrost peat plateaus already exhibit low surface moisture such that the effect of wildfire on near surface moisture and thermal conductivity is likely much more limited. These observed effects of wildfire on the seasonal cumulative ground heat flux, as indicated by end of season depths to the frost table, were strongest in the first 5-10 years following fire, and then faded and disappeared 20 to 30 years after the fire.

We also observed that wildfire has led to substantial expansion of taliks and increased soil temperatures at depth, yet these effects were delayed and were most pronounced 10 to 20 years following the fire (Fig 2.4a and Fig 2.4b). During this time period 72-100% of a burned site was underlain with a talik, compared to ~20% of unburned sites (Fig 2.4a). A similar pattern was found for maximum seasonal soil temperatures at 40cm depth for which burned sites exceeded that of the unburned sites by as much as 4°C, and ultimately reached a maximum temperature of 10.7º C, 16 years post-fire (Fig 2.4b). This is in contrast to southern permafrost free bogs where post fire, deep soils remain buffered from changes in surface soil temperatures (25). The development of taliks during this time period results from there not being enough heat loss during the winter to re-freeze the seasonally thawed layer. This is likely caused by both a greater heat loss requirement for burned sites and reduced heat loss during the winter in burned sites as a result of deeper snow due to the loss of an intercepting canopy. We interpret these delayed effects on talik coverage and maximum soil temperatures at 40cm as the cumulative multi-year effects of a perturbed surface energy balance. Following the disappearance of the short term effects (ie. increased depths to the frost table and increased thaw rates, 20-30 years post fire) is when these longer term effects start to recover (Fig 2.4a and Fig 2.5b) where by the taliks start to refreeze and soil temperatures decrease.

We observed no apparent influence of fire severity on the trajectories of the soil thermal regime following wildfire. Fire severity, as measured by depth of burn into the organic layer, has been shown to potentially affect successional trajectories in other boreal ecosystems (26). The ten burned sites varied with regards to fire size and seasonality (*Supplementary Appendix*, Table S2.1), which both can affect severity, yet neither measure influenced the observed soil thermal regime nor vegetation recovery. Given that peat plateau fires are predominantly crown fires, and in our sites fire only burned off a small fraction of the peat, we suspect there is likely little influence on the recruitment of black spruce seedlings or understory vegetation. This is important as both of these factors are known to be strongly influenced by deep organic layer removal during high severity fires (27, 28). Furthermore, we find that even a low severity fire will cause tree mortality (loss of intercepting canopy) and the loss of lichen, as indicated by the loss of tree canopy and complete lack of lichen species in sites that burned <20 years ago (Fig 2.3.). Given that we attribute loss of lichen and an intercepting canopy as the major driver of the perturbed energy balance (which results regardless of the severity of the fire) we do not expect a large influence of fire severity on the energy imbalance and subsequently increasing depths to the frost table.

The recovery of the soil thermal regime is likely linked to vegetation recovery, which strongly regulates the surface energy balance (29). Recovery of the seasonally thawed layer following fire coincides with the recovery of the lichen (*Cladonia spp.*) and black spruce trees (Fig 2.4c) Recovery of talik coverage and deep soil temperatures take an additional 10 years before they cannot be noticed at all. This is indicative that these factors represent the inter-annual cumulative effects on the energy balance. Wildfire likely affects the surface energy balance during both winter, spring, and summer. In the winter the loss of an intercepting canopy can cause a thickening of snowpack which increases insulation (30), while in the spring decreased albedo due to the loss of lichen can increase convective heat transport and thus affects snow ablation (31), and in the summer the lack of shading and decreased albedo (29) causes increased energy transfer through the soils. Taking these alterations to the energy balance into account, we created a conceptual model that identifies an important link between the vertical energy balance and lateral thaw rates that result in thermokarst bog expansion. We propose that for permafrost to persist in peat plateaus, which are surrounded by warmer unfrozen peat soils, it is necessary that there is a substantial heat loss to the atmosphere in the winter, since there is net heat gain both from below the permafrost and from surrounding non-permafrost bogs. We suspect that if an unfrozen soil layer persists, there can be no heat loss from the permafrost core to the atmosphere during the winter and this will result in a net increase in energy of the permafrost core until the soil thermal regime has recovered (i.e. 30 years). Consequently, this causes increased thaw in the form of young thermokarst bog development (Fig 2.2).

2.4.2 Effects of wildfire on young thermokarst bog expansion

Recently developed young thermokarst bogs along thawing peat plateau edges have distinct vegetation composition that is clearly discernable as bright green in high resolution satellite imagery (Fig. 2.1 and Fig. 2.5). Radiocarbon dating of thermokarst bog peat cores from the study region have shown that dominance of *Sphagnum riparium* and *Carex aquatilis* lasts for between 60 and 140 years following permafrost thaw before peat accumulation causes a succession to dominance of rust-colored *Sphagnum fuscum* and ericaceous shrubs (15) (*Supplementary Appendix,* Table S2.2). We took advantage of these ecological shifts to determine the long term effects of wildfire on permafrost thaw by assessing the spatial extents of peat plateaus, young thermokarst bogs and mature thermokarst bogs within four large peatlands (Fig 2.1). The four chosen peatlands were partially affected by wildfires 20-30 years ago (*Supplementary Appendix,* Table S2.3), i.e. the period over which field data showed that they were perturbed and thus higher rates of thermokarst bog expansion can be expected. Within each peatland, we interpreted the areal extents of young thermokarst bogs as indicators of permafrost thaw over the last 100 years, i.e. a period which includes and extends beyond the potential effects of fire.

Supervised classification using high resolution satellite imagery was carried out for 250 × 250 m peatland sections that were visually deemed to only contain peat plateaus, young and old thermokarst bogs (*Supplementary Appendix,* Fig S2.6) (see methods). At least 34, 250 × 250m cells for each burned and unburned peatlands site were classified, where the burned and the unburned peatland locations were located <20km apart. Using validation transects of the young thermokarst - mature thermokarst bog transition edge and the young thermokarst bog – peat plateau edges (*Supplementary Appendix*, Fig S2.7) we found young thermokarst bog area to generally be underestimated in the burned site and overestimated in the unburned site (*Supplementary Appendix*, Fig S2.8). Overall precision was found to be that 80% of the time the classification edges were within 1 m of the field delineated edges. This suggests that measured difference in young thermokarst bog area between the burned and unburned area is not due to difference in the classification of these two areas and we therefore deem the classification of young thermokarst to be both precise and reliable indicator of recent permafrost thaw difference between burned and unburned areas (last 100 years).

The area of young thermokarst bogs were on average ~65% greater within 20- to 30-year-old fire scars compared to nearby unburned peatland sections. Average area coverage of young thermokarst bogs were 8.8% in burned and 5.9% in the unburned areas (Fig 2.6a.). Variation in the amount of young thermokarst bog abundance among the unburned sites was likely driven by variable mean annual temperatures. The lowest young thermokarst bog abundance (~2.5% peatland coverage) was found for the Zama and Trout Lake sites, located at elevations ~300 m higher than the other sites, while the lower elevation Fort Simpson site (5% peatland coverage) is located ~250 km north of the 60th parallel site (10%). While an absolute effect of fire at +3% was similar across sites, the relative importance of fire for permafrost thaw was greater at the colder sites, compared with with 30% more young thermokarst bog in warm sites compared 250% more young thermokarst bog in colder sites.

The effect of wildfire on young thermokarst bog development rates is however, greater than indicated by current coverage of thermokarst bogs, since much of the young thermokarst bogs currently present in peatlands likely developed before fires that occurred 20-30 years ago (Fig 2.6b). Estimating differences in rates of young thermokarst development between burned and unburned peatland sections in the decades after wildfire requires a set of assumptions (*Supplementary Appendix,* Fig S2.9) of thermokarst development. In brief, we assumed that young thermokarst bogs currently present in peatlands started developing sometime between 50 and 150 years ago (15) (*Supplementary Appendix*, Table S2.2), and that it is likely that rates of thermokarst bog expansion has increased in the last 30 years in unburned peatlands due to a warming climate (18–20). Based on our assumptions, we estimate that the rate of young thermokarst bog development has nearly tripled within burned peatlands compared to nearby unburned peatlands, at 0.28 ±0.11 and 0.10 ±0.07 % yr^{-1} (95% CI), respectively (Fig 2.6b).

Development rates of young thermokarst bogs can also be expressed as rates of peat plateau loss. Across the four classified peatland sites, coverage of peat plateaus varied between 50 and 75% (*Supplementary Appendix*, Table S2.3). As such, our estimated rates of peat plateau loss in the last 30 years were 0.39 ±0.18 % yr^{-1} and 0.16 ±0.12 % yr^{-1} (95% CI), respectively for burned and unburned peatlands. Our rate of peat plateau loss due to thermokarst bog expansion within unburned areas is thus similar but in the low range of what other studies have estimated used historical image change detection (0.26-0.34 % yr^{-1} peat plateau loss) (18, 19). By combining the distribution of peat plateaus, the distribution and timing of fires (Fig 2.1), and our estimated rates of peat plateau loss within and outside burned peatlands, we estimate that the total permafrost loss over the last 30 years within the study region is 7,600 ± 4,100 km^2 (95% CI) (See supplementary information). We estimate that 2,200 ±1,500 km^2 of peat plateau thaw is directly attributed to wildfire, i.e. ~23% of the total permafrost loss. As such, wildfire has been a major driver of recent permafrost thaw in the western boreal forest of Canada.

2.5 Conclusion

In this study, we assessed the impacts of wildfire on peat plateau soil thermal regimes and young thermokarst bog expansion to provide a spatial estimate of permafrost thaw attributed to recent fires across the Taiga Plains discontinuous permafrost zone in northwestern Canada. Our findings, in conjunction with earlier research (32–34), highlight the importance of wildfire as a driver of permafrost thaw. This is particularly pressing given the known effect of climate change on wildfire activity within the boreal region (35–37). With larger fire years burning later into the season it is predicted that greater areas of permafrost soils will burn and, based on our findings this will trigger widespread thermokarst formation, as we have show that wildfire leads to a near tripling in the rate of recent thermokarst formation for ~30 years post fire. As the circumpolar regions continue to warm at twice that of the global average (38) these ecosystems will be increasingly stressed and it is unlikely that permafrost will regenerate at the rate at which we are losing it. Areas that have undergone thermokarst formation represent a potentially important player in the positive feedback loop as recently formed thermokarst features are known to have the potential to cause large methane emission (39, 40) . If we apply our estimated 25% thaw due to wildfire on methane emission from recent thermokarst features it suggests that potentially 0.055 Gg CH_4 yr^{-1} has been released to the atmosphere simply from land cover shifts associated with recent thermokarst formation just in this study region alone.

Globally these peatland permafrost features are highly significant, covering 8% of the circumpolar world and residing within the boreal forest that is subjected to wildfire every 100-120 years. Our study region is representative of boreal wetland thermokarst landscapes (41), therefore our results will help us understand global impacts of wildfire on permafrost stability.

2.6 Materials and Methods

2.6.1 Site selection for soil thermal regime monitoring

Field data for this study was collected at representative sites within the discontinuous permafrost zone of the Taiga Plains (Fig. 2.1a). Within the study area, permafrost peat plateaus comprise 33% of the landscape, storing over 24 Pg of carbon in the upper 300cm. In the past 30 years, 25% of these permafrost peat plateaus have been affected by wildfire. Six locations were chosen in areas with one unburned site and at least one historically burned site that vary in time since fire (2, 3, 4, 9, 10, 15 ,21 ,33 ,40 , 48 years ago) (Fig. 2.1, *Supplemental Appendix*, Table S2.1). Each of these sites contained at least 2 m of organic peat soil (*Supplemental Appendix*, Table S2.1). Unburned sites did not differ in mean annual air temperature (*Supplemental Appendix*, Table S2.1) and had similar vegetation composition, which consisted principally of stunted black spruce and small evergreen shrubs. Lichen and feather mosses dominated areas of less defined microtopography whereas *Sphagnum spp.* generally dominated hummocks and hollows, which covered <20% of a given site. More details in *SI Appendix*.

2.6.2 Depth to frost table monitoring

In each site 100 permanently marked points were established in a 10 x 10 grid with 3 m spacing centrally located on the peat plateau (*Supplemental Appendix,* Fig S2.1). At each point we performed repeated measures of thaw depth up to 150 cm and recorded the microtopography (hummock, hollow, undefined) and dominant ground cover (lichen, *Sphagnum spp.*, feathermoss, char). Hollows were excluded from the analysis as they are known to experience hydrological convergence which results in deeper thaw depths due to higher heat conductance (16) (*Supplemental Appendix,* Table S2.1). Depth to frost table was measured in early June, mid-July, and early-September. At the 2012, 2008, 2000 and their paired unburned site depth to frost table was measured 7 times through the summer to allow for determination of locations with taliks. More details in *SI appendix*.

2.6.3 Air and soil temperature monitoring

In two locations in each site, an Onset Hobo 8k Pendant logger was installed in a lichen/char covered area at 10cm and 40cm depth. An additional logger was placed under a white shade on a pole ~1m in height at each site to monitor air temperatures. Loggers recorded temperatures every 2 hours for a year from September 2015 – September 2016, and were then aggregated for a daily average. More details in *SI appendix*.

2.6.4 Vegetation Surveys

During August 2016, vegetation analysis was completed using the point intercept method (42) along three 30 m transects in each site. Tree heights of all black spruce trees within a meter on each side of the transects was visually estimated. More details in *SI appendix*.

2.6.5 Image acquisition and preparation for supervised peatland classification

Multispectral WorldView2 images were obtained for four sites within the study area (Fig 2.1a). All images were taken between June and September in 2011. At each location, two 5×5 km images were acquired, one including burned areas and one adjacent unburned area. Within the acquired 5×5 km satellite image there were substantial areas that were unsuitable for analysis as they contained fens, uplands, and lakes. To examine the influence of wildfire on permafrost stability solely in peatlands, 500 sections (250 × 250 m) were formed from the original 5x5 km image (*Supplementary Appendix*, Fig S2.6). Cells that contained only peat plateaus, young and mature thermokarst bogs were retained for further analysis. This resulted in 57-61 cells, depending on the location of image (*Supplementary Appendix,* Table S2.3). More details in *SI appendix*.

2.6.6 Supervised peatland classification

Using our field knowledge we defined three spectrally separable land covers types: peat plateaus, young thermokarst bog, and mature thermokarst (Fig 2.5). Maximum likelihood classification was performed in ENVI based on the spectral signatures of these three classes. More details in *SI appendix*.

2.6.7 Accuracy assessment of peatland classification

Classification validation transects were collected in August 2016 at the Zama burned and Zama unburned location using a dGPS unit accurate to 10cm. The user defined edge of the young thermokarst bog to mature thermokarst bog edge and the young thermokarst bog to peat plateau edge were walked and GPS points collected approximately every 2 meters (*Supplementary Appendix* , Fig S2.7). To assess classification accuracy the distance between the user collected field edge and the classification edge was determined (*Supplementary Appendix,* Fig S2.8.).

2.6.8 Estimating trajectories of young thermokarst development

A range of possible trajectories based on timing of recent thermokarst bog development, percent thaw per year as a result of fire, at each site were estimated using a boot strapping approach (~5000 iterations). Such estimations require a set of assumptions detailed in *SI appendix*.

2.6.9 Spatial scaling of peat plateau loss due to thermokarst development

To scale peat plateau thaw to the scale of the study region we applied our estimated rate of young thermokarst development per year to the total burn area within the study area. More details in *SI Appendix*.

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2. 9 SI Appendix

2.9.1 Site selection for soil thermal regime monitoring

The Northern Circumpolar Soil Carbon Database (version 2; (1)) was used to determine the area of permafrost peat plateaus within the study area. The database provides spatially distributed quantification of soil organic carbon at 0-300 cm depths and soil type across the circumpolar permafrost region. Burned area from the past 30 years from the Canadian National Fire Data (2) base was overlaid. Proportion histel burned was determined by multiplying area of burned polygon by proportion of histels within given area.

2.9.2 Thaw depth monitoring

A Kruskal-Wallace one way analysis of variance for non-parametric data was performed to identify difference between points within a grid at a given location based on landcover and topography. At all site hollows displayed significantly deeper depths to frost table ($p<0.05$), regardless of landcover and were therefore excluded from analysis.

Depth to frost table was determined using a kernel density estimator - a nonparametric technique way to estimate the probability density function of a random variable (3) as it is more representative of the ecological dynamics than mean or median (*Supplementary Appendix,* Fig S4). By using the max density estimator (4) we were able to identify locations where there are no underground hydrological convergence points (5). A kernel density estimation allowed us to more accurately represent the depth to frost table in locations where taliks are absent, as using either a mean or median thaw depth would strongly influence our estimation of active layer depth based on talik frequency.

Talik distribution was determined using end of thawing season (September) measurements of depth to frost table. Locations with thaw depths greater than 100cm were deemed to be underlain with a talik (Fig S2.3). To verify this we performed an error matrix with sites where thaw depths had been more frequently monitored. From May to July, thaw depths in the 2012 burn sites and its paired unburn site were measured four times over the course of the early summer (May-July). We manually identified locations of taliks as places where thaw depths doubled between two consecutive sampling periods. We then set an approximation that end of season thaw depths that were 90cm ± 40% would have a talik. We tested this against our manually identified locations of taliks and performed an error matrix. Our accuracy assessment for predictions of absence or presence of taliks within a burned peat plateau site using September thaw depths showed 92% accuracy in the unburned site and 97% in the burned site (Table S2.4 and Table S2.5) .

2.9.3 Air and soil temperature monitoring

Soil temperature loggers were installed in areas of undefined topography in lichen or char to avoid areas of increased conduction due to hollows or the presence of *Sphagnum spp.* Soil temperatures were averaged between the two loggers in each site. During time periods where there was logger failure, logger data from the remaining functioning logger only was used.

2.9.4 Vegetation Surveys

Vegetation surveys were completed by delineating a 50 m baseline transect along a randomly chosen geographic bearing. Three 30 m transects were established perpendicular to the 50m transect: one at the center at the 25 m mark and two on either side of the center at randomly generated distances greater than 5 m away from the center point. Understory species abundance was measured using the point intercept method (6) along all three transects with a pin being dropped every meter. Species abundance was calculated using the following formula: Species abundance (% cover) = (number of hits per species/total number of hits)*100. Tree heights were visually estimated within 1 m of either side of each 30 m transect for a total sampling area of $180m^2$.

2.9.5 Image acquisition and preparation for supervised peatland classification

The commercial WorldView2 satellite provides 46 cm panchromatic resolution with three visible bands within the visible spectrum (blue 450 - 520 nm; green 520 - 600 nm; red 625 - 695 nm).

2.9.6 *Supervised peatland classification*

Within the WorldView2 imagery plateaus are characterized by relatively high albedo *Cladonia* spp., speckled with relatively low albedo black spruce. Bogs are characterized by spectrally bright *Sphagnum* spp. and thermokarst by spectrally bright *Sphagnum riparium* and graminoids (Fig 2.5). Post classification sieve and clumping was done to remove small numbers of pixels within larger contiguous classed and to achieve a smoother overall classification.

2.9.7 Trajectories of young thermokarst bog development

A range of possible trajectories of young thermokarst bog development at each site were estimated with the following assumptions of the boot strap exercise. 1) Sections that burned 30 years ago had identical trajectory as unburned sections up until the time of fire. 2) The young thermokarst bogs that we see today on the landscape started developing 100 ± 50 years ago (95% CI), as indicated by ^{14}C and Pb dating of peat cores (Table S2.2). 3) We assume that the upper limit of the 95% CI range for young thermokarst bog development rate before the fire event would be similar as if there has been no increase in rate in the last 30 years due to a warming climate. However, ample evidence exists to suggest that thermokarst bog expansion have increased as a result of climate warming. We therefore deem it highly unlikely that that thermokarst expansion rates in unburned sections were higher prior to the time of the fire as compared to after. We define this rate as current median thermokarst extent in unburned sections divided by years since currently present thermokarst bogs started developing (i.e. age from distribution described in assumption 2). 4) We assume that the lower limit of the 95% CI range for young thermokarst bog development rate before the fire is half of the rate calculated from the upper limit 95% CI development rate before the fire, i.e. current median thermokarst extent in unburned sections divided by years since currently present thermokarst bogs started developing divided by 2.

Simply, it is unlikely that there had been no thermokarst development in the period before the fire. It is also unlikely that the rates of young thermokarst bog development were higher in the period before the fire compared to after for unburned sections. Our central estimate of the amount of young thermokarst bog on the landscape at the time of the fire is half of present day coverage, which is consistent with several studies that suggest a doubling in the rate of thermokarst development in the past 30 years (7–9). Following the fire event, we assume a constant rate of young bog development that led to currently present observed young thermokarst bog coverage (Fig S2.9).

This analysis was completed for each location and the average and standard deviation of mean development rates across the four sites was used to create our estimated trajectory of young thermokarst development (Fig 2.6b).

2.9.8 Scaling of regional peat plateau loss due to thermokarst expansion

Using the Northern Circumpolar Soil Carbon Database (version 2; (1)) and the Canadian National Fire Data (1) we determined the area of histel burned each year for the past 60 years. We assume fire affected areas thaw at an increased rate for 30 years. Therefore, observed present day young thermokarst bogs are the result of fires that occurred prior to 1980s whose effects are beginning fade and transition to mature thermokarst, and post 1980s whose effects are only beginning to be seen. We applied the increased fire-induced rate of thaw to the histel area burned within a given year times the number of years since fire, or pre-30 years ago. From this derived total thaw area due to fire. Total thaw area due to climate warming alone was derived by multiplying total histel area within study area by our estimate of thaw from the unburned sites. Our central estimate of thaw rate and standard deviation was determined using a bootstrapping approach (~300 iterations) to determine fire-induced thaw area and climate induced thaw area.

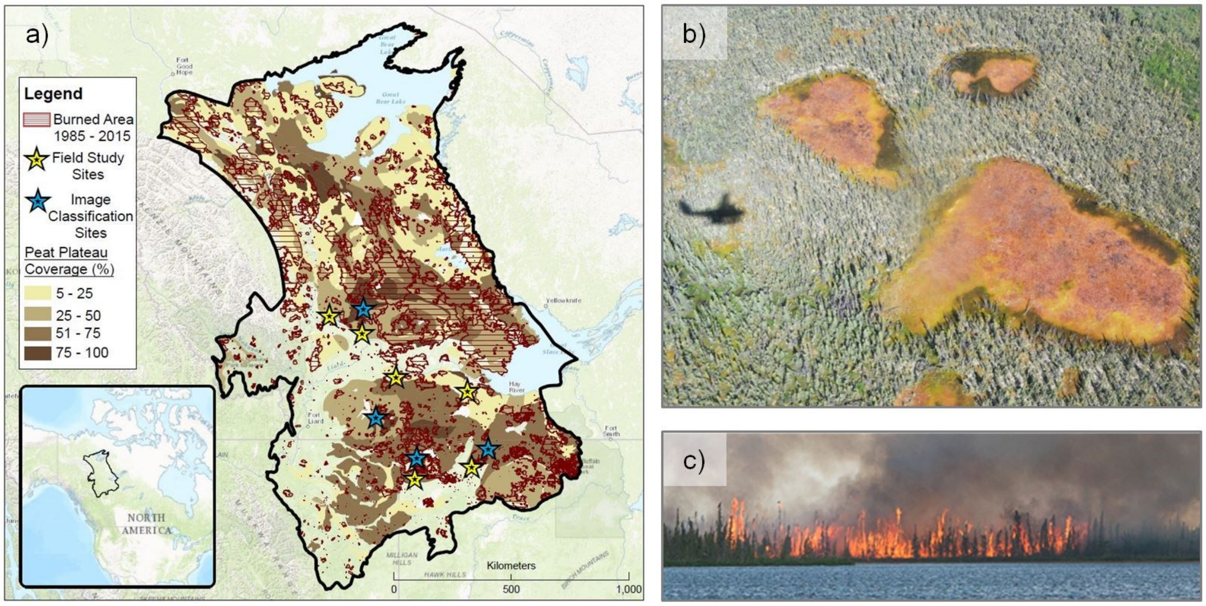

Fig. 2.1. Wildfire in western Canadian boreal peatlands. A) Outline of the study region, defined as the discontinuous permafrost zone (21) within the Taiga Plains ecozone (22). Shading within the study region shows the coverage of peat plateaus as indicated by histel soils (23), while historical burned areas from the last 30 years are shown as hashed areas (24). Yellow stars indicate locations where field data characterizing soil thermal regimes were collected, with one unburned and at least one burned site monitored at each star location. Blue stars indicate locations where satellite image classification of recent thermokarst bog expansion was carried out within extensive peatlands partially affected by wildfire. B) Example of a peatland characteristic for the study region, with three open non-permafrost thermokarst bogs surrounded by treed permafrost-affected peat plateaus. Distinct young thermokarst bog stage is visible as green areas along some peat plateau edges, a result of recent permafrost thaw. Photo copyright Mason Stothart. C) Example of a characteristic crown fire burning a peat plateau, near Fort Simpson, NWT, in 2013. Photo copy right Franco Alo.

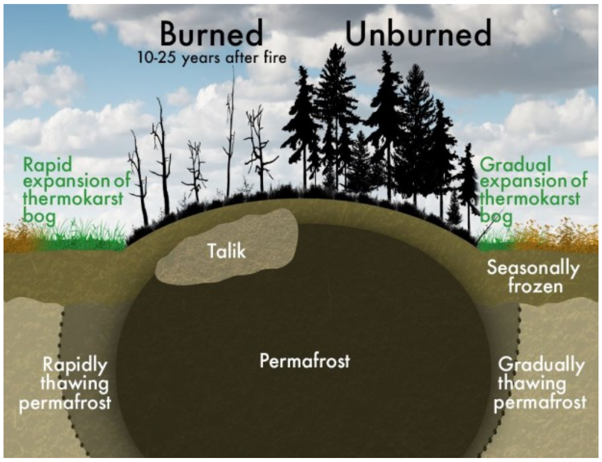

Fig. 2.2. Simplified illustration of soil thermal states for peat profiles along transects from thermokarst bogs to peat plateaus within and outside historical burned areas.

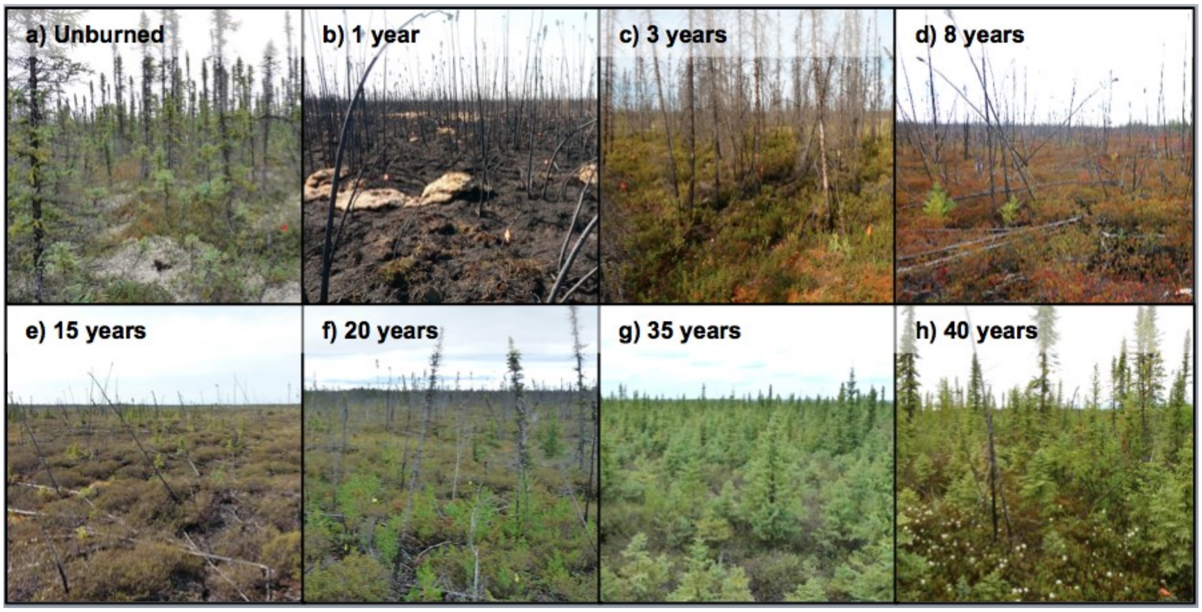

Fig. 2.3. Vegetation recovery at select peat plateau sites used in this study to assess long term effects of fire on soil thermal regime.

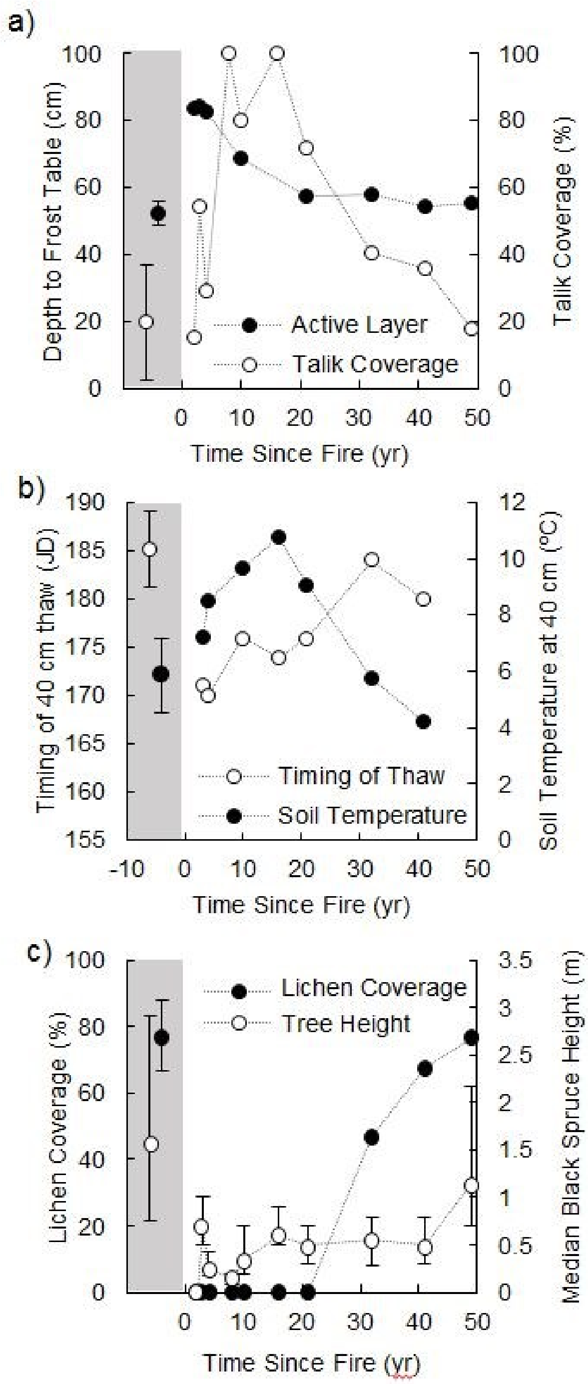

Fig. 2.4. Trajectories of soil thermal regime and vegetation recovery on peat plateaus following wildfire. A) Trajectories of depth to frost table at point without taliks, and proportion of points where thaw depth thaw monitored that had continuously thawed soil layers, i.e. taliks. No depths to frost table estimates were possible for sites burned 7 and 16 years ago due to 100% talik coverage. B) Trajectories rate of that, as indicated by soils at 40cm rising above 0.5ºC, and maximum seasonal soil temperature attained at 40 cm depth (points at 7 and 49 years post fire not available). C) Trajectories of vegetation recovery, exemplified by the percent ground cover of lichens, and median height of black spruce trees. In each panel, the average of 6 unburned sites is shown on the left with grey background, where error bars indicate ±1 standard deviation for percent lichen and interquartile range for black spruce trees.

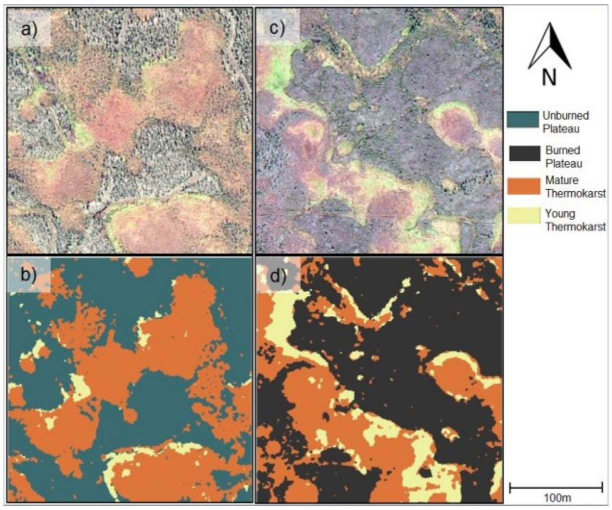

Fig 2.5. Classification of peat plateau, young thermokarst bog, and mature thermokarst bog extents in large peatlands using high resolution satellite imagery (World View 2, 50 cm resolution). A) and C) Supervised classification was carried out for 250 x 250 m sections of unburned and burned peatlands respectively that were visually assessed to only contain peat plateau, young thermokarst bog, and mature thermokarst bogs. Sections containing channel fens, upland forests, ponds, clouds, or shadows were excluded from analysis and are shown with grey shading. B) and D) Example of a 250 × 250 m section selected for classification in the unburned and burned peatlands respectively.

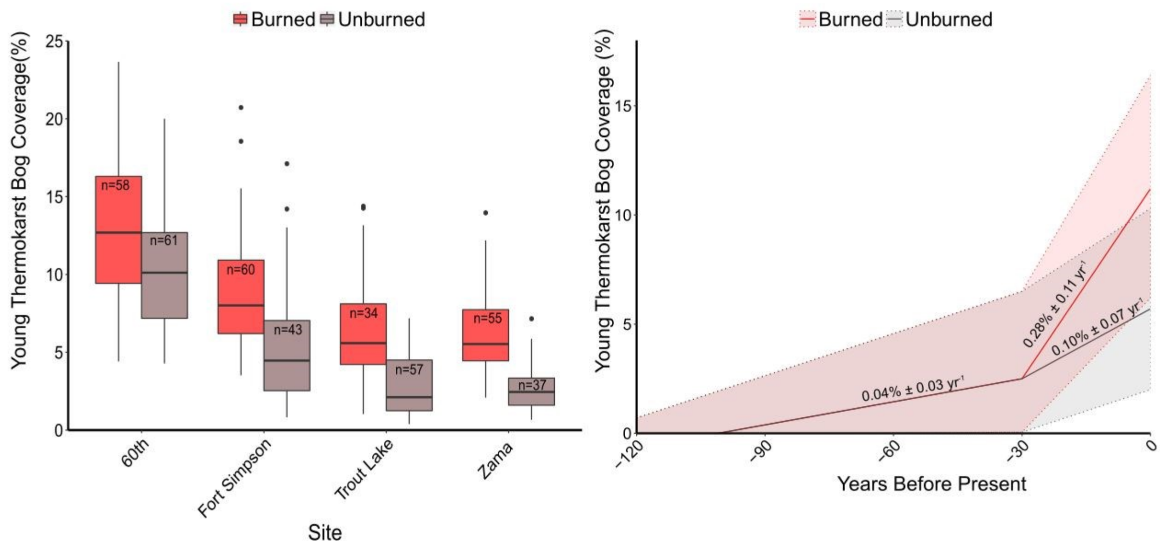

Figure 2.6. Effect of wildfire on permafrost thaw through development of young thermokarst bogs in boreal western Canada. A) Current-day coverage of young thermokarst bogs within and outside 20- to 30-year-old fire scars at four large peatlands. Box plots indicate median, interquartile range, and minimum and maximum young thermokarst bog coverage among classified 250 × 250 m sections, and number of classified sections is indicated (n). In each pair, burned sections had significantly greater was significantly different from its unburned pair (*p<0.01,* pairwise t-test). Sites are ordered left to right by decreasing mean annual air temperature. B) Temporal development (± 95% CI) of young thermokarst bogs currently present inside and outside 30-year-old fire scars. Young thermokarst bogs persist in peatlands for 100 ± 40 years before succession into mature thermokarst bogs (15), suggesting that a significant proportion of the currently present young thermokarst bogs formed prior to the historical wildfire. Rate of development of young thermokarst bog, expressed as a % of total peatland area (± 95% CI), shown above each trajectory.

SI Figures and Tables

Table S2.1: Summary statistics from chronosequence sites. Date and size of fire derived from Large Fire Data Base (NRCAN 2016). Number of active layer measurements include only locations used in analysis, ie. hollows excluded.

Site	**Site Coordinates**	**Elevation (masl)**	**Date of Fire**	**Size of Fire (ha)**	**Peat Depth (cm)**	**Thaw depth point locations (n)**	**Mean Annual Air Temp[1] (ºC)**
Samba Deh 1967	61°21'18"N 121° 0'29"W	237	June 20, 1967	605	>250	100	-2.89
Camsell 1975	62°17'59"N 122°38'8"W	293	June 24, 1975	32,087	230	92	-2.34
Zama 1984	59°20'45"N 119°20'24"W	587	July, 10 1982	6132	260	92	N/A[2]
Fort Simpson 1995	61°59'45"N 121°23'45"W	186	August 18,1995	4653	>300	100	-0.15
Lutose 2000	59°30'0"N 117°12'17"W	318	June 28, 2000	1471	280	100	-0.58
Fort Simpson 2006	61°59'24"N 121°23'51"W	185	July 22, 2006	808	200	100	-0.14
Lutose 2007	59°25'2"N 117°17'4"W	324	June 13, 2007	2245	>300	100	-0.91
Lutose 2012	59°46'42"N 117° 3'40"W	299	July 9, 2012	37,244	>300	100	N/A[2]
Samba Deh 2013	61°11'42"N 120° 5'25"W	290	June 27, 2013	96,330	290	99	-1.82
Kakisa 2014	60°56'26"N 117°21'42"W	225	May 26, 2014	80,311	255	100	-1.91
Camsel Unburn	62°14'46"N 122°34'23"W	262	-	-	130	99	-2.51
Samba Deh Unburn	61°11'59"N 120° 6'50"W	287	-	-	290	98	-2.04
Lutose Unburn	59°29'4"N 117°10'41"W	305	-	-	>300	95	-1.45
Zama Unburn	59°22'9"N 119°19'16"W	584	-	-	260	100	-3.1
Kakisa Unburn	61° 4'53"N 117°37'23"W	250	-	-	200	82	-1.57
Fort Simpson Unburn	61°59'34"N 121°22'59"W	186	-	-	>300	96	-0.81

[1]July 2015 – 2016

[2]Temperature sensor malfunctioned

Table S2.2. Dating of transitions between plateau peat and young thermokarst bog peat in cores collected from current young and mature thermokarst bog sites. All dating is done ~1-2 cm below the transition, which means that true transition dates happened more recently than indicated by the dating techniques. Note that these cores were not collected at the transition between young and mature thermokarst bogs, but rather centrally within young and mature thermokarst bogs, respectively, and thus in order to assess the persistence of young thermokarst bogs in the landscape, we are interested in seeing the maximum age of transitions within current young thermokarst bogs and the minimum age of transition in current mature thermokarst bogs.

Core type and ID	**Location**	**Depth of transition (cm)**	**Age of transition (cal YBP)**	**Transition dating method**	**Reference**
<u>*Young thermokarst bog cores*</u>					
SC1-E	61°18'N 121°18'W	0	25	Aerial photo	Pelletier et al. 2017
SC1-YB	61°18'N 121°18'W	61	130	^{210}Pb	Pelletier et al. 2017
SC2-YB	61°18'N 121°18'W	60	125	^{210}Pb (Model)	Olefeldt et al. in prep
SC3-E	61°18'N 121°18'W	29	60	^{210}Pb (Model)	Olefeldt et al. in prep
SC3-YB	61°18'N 121°18'W	55	105	^{210}Pb (Model)	Olefeldt et al. in prep
SC4-E	61°18'N 121°18'W	47	145	^{210}Pb	Olefeldt et al. in prep
SC4-YB	61°18'N 121°18'W	62	120	^{210}Pb	Olefeldt et al. in prep
L1-YB	59°28'N 117°10'W	64	130	^{210}Pb (Model)	Heffernan et al. in prep
L1-IB	59°28'N 117°10'W	42	76	^{14}C	Heffernan et al. in prep
Maximum age of transition			*145*		
<u>*Mature thermokarst bog cores*</u>					
SC1-OB	61°18'N 121°18'W	130	530	^{14}C	Pelletier et al. 2017
SC2-OB	61°18'N 121°18'W	89	367	^{14}C	Olefeldt et al. in prep
SC3-OB	61°18'N 121°18'W	66	143	^{14}C	Olefeldt et al. in prep
SC4-OB	61°18'N 121°18'W	120	587	^{14}C	Olefeldt et al. in prep
L1-OB	59°28'N 117°10'W	89	362	^{14}C	Heffernan et al. in prep
Minimum age of transition			*143*		

Table S3.3. Summary statistics from peatland classification showing sites with site coordinates, number of 250 x 250m sections classified and percent coverage of the three peatlands classes ± one standard deviation.

Site	**Site Coordinates**	**Date of Fire (size, ha)**	**Elevation (masl)**	**Classified sections (n)**	**Peat Plateau coverage (%)**	**Young Thermokarst Bog coverage (%)**	**Mature Thermokarst Bog coverage (%)**
Zama Unburned	59°20'N 119°19'W	July 10, 1984 (6132)	584	55	74.3 ± 9.2	2.7 ± 1.3	23.0 ± 8.5
Zama Burned				37	66.8 ± 10.0	6.2 ± 2.8	26.7 ± 9.1
Trout Lake Unburned	60°26'N 120°51'W	June 28, 1996 (13,202)	559	34	70.3 ± 13.3	2.7 ± 1.9	23.9 ± 12
Trout Lake Burned				57	65.52 ± 8.4	6.3 ± 3.3	27.8 ± 8.2
Fort Simpson Unburned	62°25'N 121°20'W	August 3, 1995 (238,566)	260	60	49.4 ± 10.7	5.1 ± 3.4	44.7 ± 11.8
Fort Simpson Burned				43	52.9 ± 11.1	9.0 ± 3.9	37.7 ± 9.0
Sixtieth Burned	59°51'N 116°33'W	July 12, 1981 (552,539)	314	61	65.8 ± 8.4	13.0 ± 4.8	21.1 ± 8.0
Sixtieth Unburned				58	55.0 ± 10.0	10.5 ± 3.9	34.3 ± 9.9

Table S2.4. Accuracy assessment for predictions of absence or presence of taliks within a burned peat plateau site using September thaw depths. Data is from one unburned site, Lutose Unburned (See table S2.1), where actual absence or presence of taliks was determined by seasonal monitoring of thaw depths.

	Predicted Absence of Talik	**Predicted Presence of Talik**	**Sum**
Actual Absence of Talik	63	4	67
Actual Presence of Talik	4	24	28
Sum	67	28	95

Overall accuracy = 92%

Table S2.5. Accuracy assessment for predictions of absence or presence of taliks within a burned peat plateau site using September thaw depths. Data is from one burned site, Lutose 2012 (See table S2.1), where actual absence or presence of taliks was determined by frequent monitoring of thaw depths.

	Predicted Absence of Talik	**Predicted Presence of Talik**	**Sum**
Actual Absence of Talik	36	2	38
Actual Presence of Talik	1	55	56
Sum	37	57	95

Overall accuracy = 97%

Figure S2.1. Example of a peat plateau site (Kakisa 2014) where soil thermal regime was monitored. Each peat plateau site had a 5 m spaced grid with 100 point locations marked with flags where depth to frost table was measured.

Fig. S2.2. Example of transects from peat plateau, through young thermokarst bog, to mature thermokarst bogs. (a) Lutose 2012, b) Samba deh 2013, c) Fort Simpson unburned, d) Lutose unburned. Notice the shift in dominant vegetation between the young and mature thermokarst bogs.

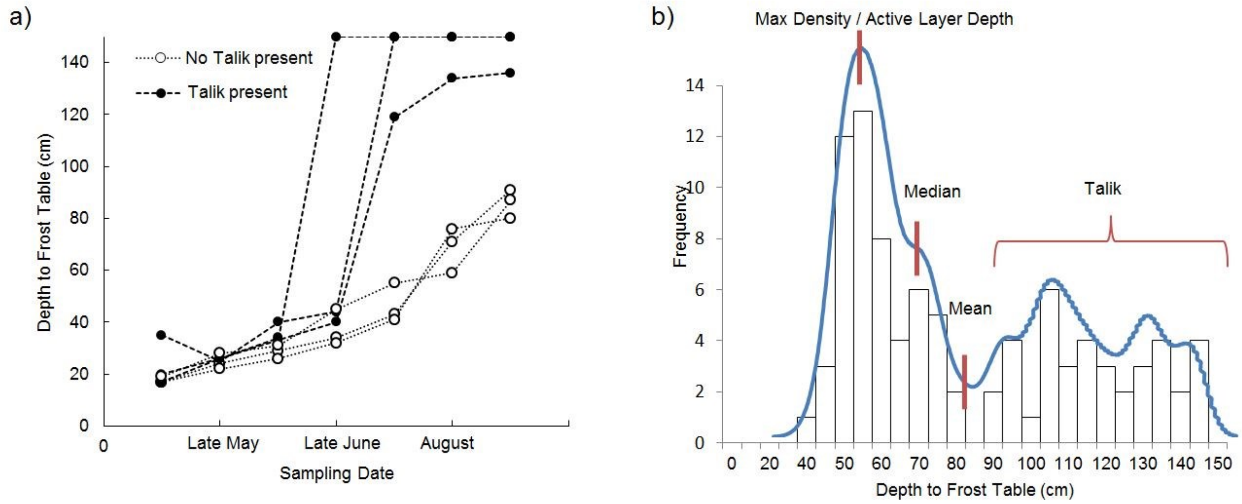

Fig S2.3. Determination of talik presence, and estimation of thaw depth at point locations where taliks were absent. A) Example of seasonal development of the thawed soil layer at 6 representative point locations at the Lutose 2012 site. Soil probe was 150 cm long, hence measures of 150 cm indicate that no frost table was reached. We classified point locations as having taliks if there was a >90 cm increase in thaw depth between two monitoring occasions, as exemplified by three point locations. Only a subset of sites had thaw depths monitored as frequently as Lutose 2012, and thus for other sites we did not have direct observations of talik frequency. B) Histogram of September point location depth to frost table measurements (n = 100) from the Fort Simpson 1995 site overlaid by a gaussian kernel density estimation (Wessa 2015). All point locations with September thaw depths >100 cm were assumed to have taliks, an assumption that was tested against observed talik presence at a few sites (See table S2 and S3). We used the maximum density to estimate representative active layer depth of point locations where taliks were absent, as using either the mean or median thaw depths would be strongly influenced by talik frequency.

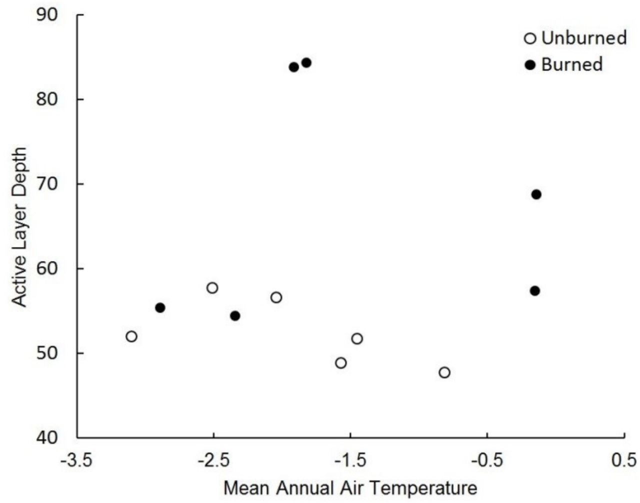

Fig S2.4. September active layer depth with mean annual air temperature in the burned and unburned sites. Climate was not found to be the main influence on active layer depth, supporting our conclusion that wildfire is the main driver of difference in active layer depth.

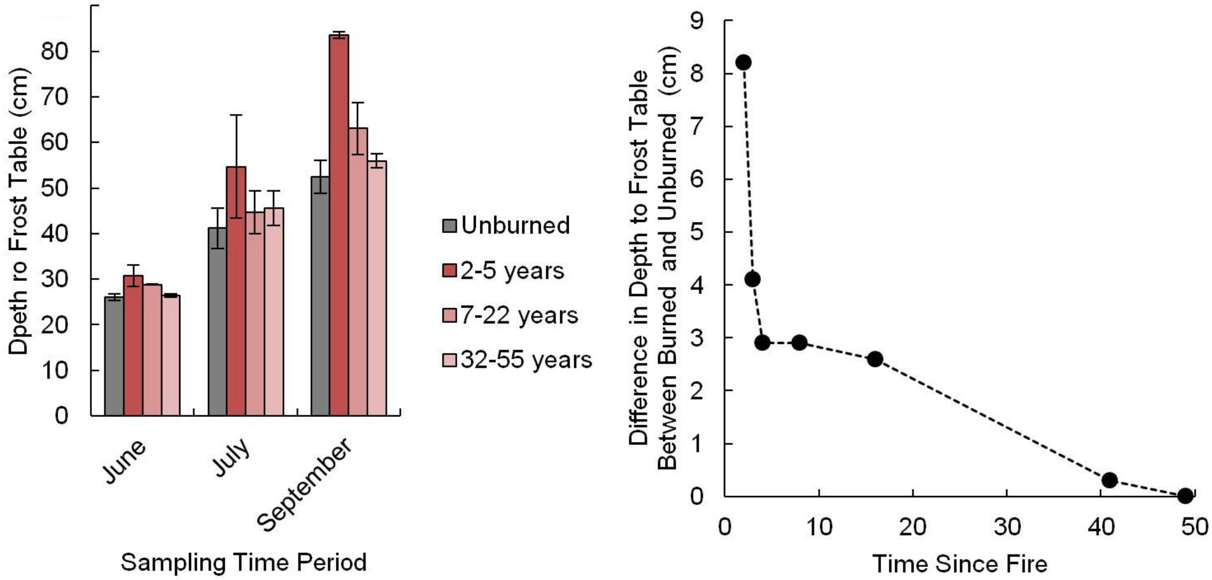

Fig S2.5. Differences in the seasonal development of the thawed soil layer between unburned and historically burned sites. A) Thaw depths are shown ±1 standard deviation of n=3 (2-5 years), n=4 (7-22 years), and n=3 (32-55 years). Point locations where taliks later developed were included for June and July dates. B) Absolute difference in depth to frost table between burned and paired unburned sites in early June. Sites 10, 21, 32 post fire not included due to inaccessibility at this time of year.

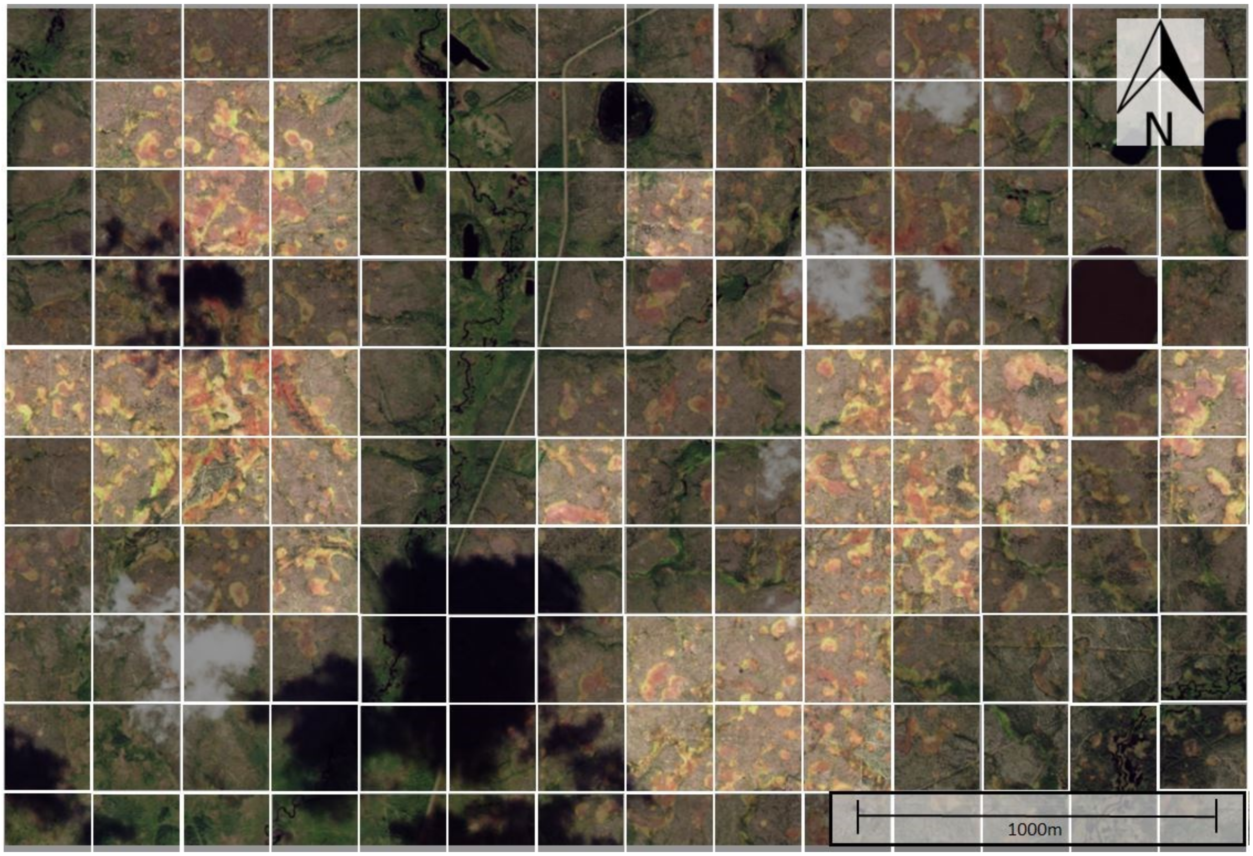

Fig S2.6. Example section of World View 2 satellite image. Supervised classification was carried out for 250 x 250 m sections of peatlands that were visually assessed to only contain peat plateau, young thermokarst bog, and mature thermokarst bogs. Sections containing channel fens, upland forests, ponds, clouds, or shadows were excluded from analysis and are shown with grey shading.

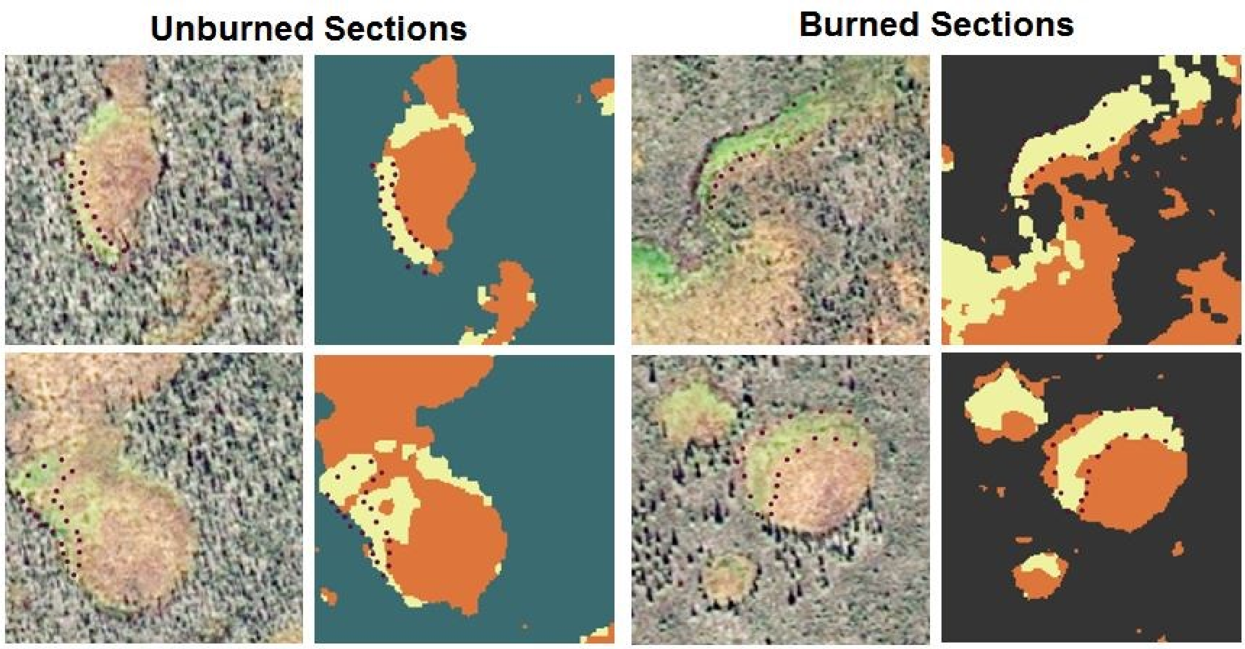

Fig S2.7. Field validation of extents of young thermokarst bog. World View 2 Satellite Image (75 × 75 m) shown right with corresponding classification on the left. Overlain points taken with dGPS unit, up to 10 cm accuracy.

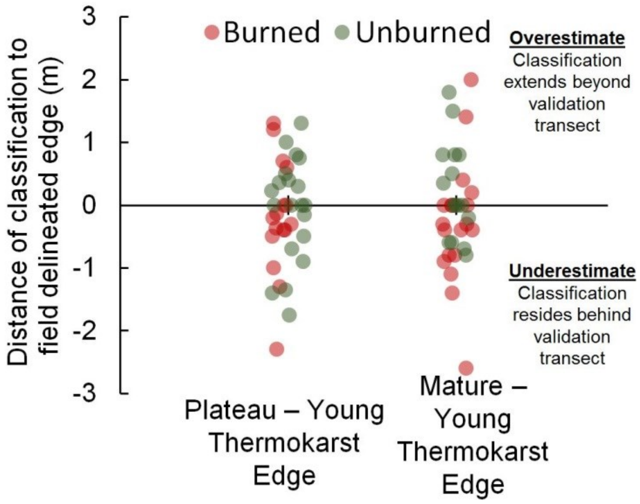

Fig. S2.8. Residual difference between edges of supervised classification and field measurements of edge. Overestimates occurred when classification is outside the field delineated young thermokarst bog area, underestimates occurred when the classification was within the field delineated young thermokarst bog area.

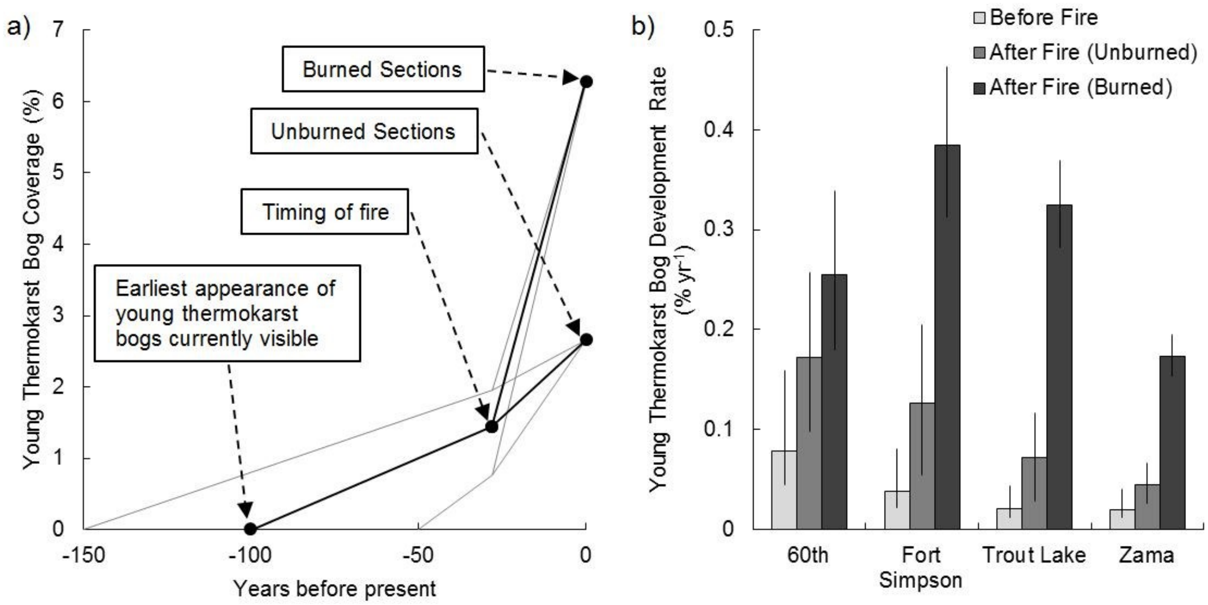

Figure S2.9. Estimate of the trajectories of young thermokarst bog development. A) Trajectories of recent thermokarst bog development based on assumptions 1-4. B) 95% confidence intervals for each site, from which mean and standard deviation of the mean were used for final trajectory estimate.

2.9.10 Supplementary Literature Cited:

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Chapter 3: Effects of wildfire on soil respiration in permafrost peatlands of the Canadian Western Boreal Forest

3.1 Abstract

Wildfire in northern boreal peatlands is contributing to widespread permafrost thaw by deepening the active layer, causing significant amounts of previously frozen soil carbon to be exposed to microbial decomposition. This may result in the release of old carbon sources to the atmosphere in the form of the greenhouse gas CO_2 but the rate and duration of such release is poorly quantified. We used field-based measurements of soil respiration in combination with a two-depth soil temperature model in burned and unburned sites to determine carbon loss due to fire-induced active layer deepening. We show that respiration during the growing season is significantly lower in burned sites, despite having active layers over 100cm deeper than unburned sites. However, we estimate the proportion of carbon respired from depth to be 40% in the burned site, which includes previously frozen carbon stores, compared to 10% in the unburned site. This results in ~32 g Cm^{-2} respired from depth in the burned site and ~8 g Cm^{-2} was loss from the unburn site. While CO_2 release from soil respiration is comparable or lower in the burned than in the unburned peatlands, likely due to the removal of the top labile peat layer during fire, fire-induced active layer deepening mobilizes carbon release from these ecosystems.

3.2 Introduction

Northern boreal forests in the discontinuous permafrost zone are experiencing drastic warming and increased fire frequencies [*Kasischke and Turetsky*, 2006] that is expected to alter the carbon cycling in these regions [*Schuur et al.*, 2015]. These altered conditions are likely to result in changes in soil temperatures and permafrost stability [*Genet et al.*, 2013; *Zhang et al.*, 2015; *Helbig et al.*, 2016] that control greenhouse gas fluxes to the atmosphere. Northern permafrost soils hold large quantities of carbon [*Hugelius et al.*, 2014] that is considered locked as it undergoes minimal microbial activity while frozen. However, deeper seasonal thaw layers as the result of permafrost thaw [*Viereck et al.*, 2008; *Brown et al.*, 2015] and thermokarst formation [*Olefeldt et al.*, 2016] expose previously frozen soil organic carbon to decomposition, potentially incorporating it into current carbon cycling. The magnitude, timing, and form of the permafrost carbon feedback remains highly uncertain [*Schuur et al.*, 2015] and increasing prevalence of disturbances such a wildfire on the landscape are furthering our uncertainty about potentially large positive feedbacks from these systems [*Schaefer et al.*, 2012; *Treat et al.*, 2013; *Koven et al.*, 2015].

Wildfire is the major stand-renewing agent in these systems and is an integral part in determining patterns of vegetation and forest structure [*Rowe and Scotter*, 1973]. Wildfire within peatlands represent a large source of carbon to atmosphere, burning over 1500km^2 and releasing 6TgC annually in Western Canada alone. Wildfire affects forested permafrost peat plateaus in similar frequencies to upland forests (100-120 year fire return interval; [*Turetsky et al.*, 2004]) and in the last 30 years more than 25% of permafrost peatlands within the Taiga Plains ecozone have burned [*Gibson et al. in prep*]. Climate change is likely to exacerbate the effect of fire on carbon emissions from peatlands due to changes in the fire regime and drought frequency and intensity [*Flannigan et al.*, 2009]. Across North America the total annual burn area has increased due to higher mean annual temperatures and drier conditions [*Kasischke and Turetsky*, 2006] and is predicted to increase by 25% by the year 2030 and the 75-100% by the end of the century [*Wotton et al.*, 2010].

The increase in fire prevalence in these permafrost landscapes is having significant impacts on permafrost stability, and therefore the stability of these soil organic carbon reserves. Studies have shown that fire can increase soil temperatures for several decades and lead to thickening of the active layer [*Mackay*, 1995; *Burn*, 1998; *Yoshikawa et al.*, 2002; *Viereck et al.*, 2008; *Rocha et al.*, 2012; *Nossov et al.*, 2013; *Brown et al.*, 2015, *Gibson et al. in prep*]. Despite the rapid recovery of understory vegetation that has been shown to dramatically increase surface albedo [*Randerson et al.*, 2006; *Amiro et al.*, 2009], the most pronounced effect of wildfire on soil temperatures in permafrost peatlands occurs 10-20 years post fire where soils temperatures at depth can be as much as 4°C warmer [*Gibson et al. in prep.*]. Many of the observations on active layer thickening come from Alaskan upland systems or lowlands with shallow peat layers (<1m [*Viereck et al.*, 2008; *Brown et al.*, 2015]) where they show active layer deepening of 20-50cm [*Treat et al.*, 2013]. However, in the permafrost peatlands of Western Canada where peat depths range from 2 to 6 m, active layers can increase up to 100cm in depth after a fire [*Gibson et al. in prep.*]. This combination of increased soil temperatures and increased seasonally thawed soils would tend towards an increase in net carbon loss from post-fire permafrost soil. However, we have very few measurements on this phenomenon.

Much of our current understanding of the response of soil respiration from thawed permafrost soils comes from incubations experiments. Soil microbial respiration generally increases exponentially with temperature [*Davidson and Janssens*, 2006] and Schadel et al. (2016), using meta-analysis from 21 incubation studies, showed that increasing soil temperatures led to a net increase in carbon release by a factor of 2.0. Furthermore, the amount of carbon released under aerobic conditions was on average 3.4 times higher than anaerobic incubations conditions. This is particularly important in post fire peat plateau systems where permafrost thaw leads to a deepening and warming of the aerobic layer, which could drive the mobilization of stored older carbon.

The temperature dependence of soil respiration may be modeled by a constant Q_{10} which is the proportional increase in the rate of respiration for a change of 10°C and is well established to range between 2 and 2.5 [eg. *Knorr et al.*, 2005]. Most models use near surface temperatures which are found to be relatively good predictors of ecosystem respiration [*Lafleur et al.*, 2005]. This is likely because a large portion of CO_2 released comes from near surface sources associated with vegetation and the top soil layer which experiences the largest variation in temperature and represents the greatest fraction of labile carbon in the soil. However, such near-surface temperature models do not seem suitable to account for the respiration at depth when the soil thermal regime at depth is altered, particularly as a consequence of fire-induced active layer deepening. To our knowledge no study has used two temperatures (both a shallow and deep temperature, or a temperature profile) to model soil respiration within permafrost landscapes. This method would allow us to isolate contribution of soil respiration from shallow and deep old carbon sources. Given there is generally a higher temperature sensitivity of soil respiration at low temperatures [*Kirschbaum*, 1995; *Goulden*, 1998], soil organic carbon in cold regions may be highly sensitive to above-average temperatures that occur as a result of wildfire.

Previous work by Schuur et al. (2009) found that following permafrost thaw with thermokarst development up to 15% of ecosystem respiration is derived from the decomposition of old carbon. Yet presently, there remains little understanding of contribution from deep, old carbon to soil respiration following active layer deepening, particularly following a wildfire where soil temperatures at depth have been shown to be much greater. In order to address this we use *in situ* measurements to examine the effect of wildfire on post fire soil respiration from sites that burned 10-20 years ago and a paired unburn site to (1) quantify and compare soil respiration from burned and unburned permafrost peatlands, (2) determine if soil respiration models that include both a shallow and a deep temperature more accurately determine carbon loss from post fire sites, (3) and examine the contribution of soil respiration from deep, fire-induced thawed soils to total soil respiration within a site.

3.3 Methods

3.3.1 Study Area

We conducted our study within a burned and paired peatlands near Lutose, AB (59°29'4N, 117°10'43W) within the discontinuous permafrost zone of the Taiga Plains (Figure 3.1). We selected sites that burned in 2000 and 2007 and a paired unburned site (Figure 3.1). Peatlands in this area are a mosaic of permafrost plateaus, channel fens, and ombrotrophic flat bogs. Unlike surrounding bogs and fens, the permafrost peat plateaus support black spruce (*Picea mariana*) and are underlain by ~5-10m thick permafrost. The mean annual air temperature during our study period was -1.5ºC. Vegetation within the burn sites is limited to *Rhododendron groenlandicum* and <1m tall black spruce while the unburned site is dominated by black spruce 2.3 ± 1.9m tall and an understory of predominantly *Rhododendron groenlandicum, Vaccinimum vitis-idaea,* and *Vaccinium uliginosum, Cladonia* spp. and *Sphagnum* spp. Both burn sites show little to no nonvascular recovery with 95-100% char cover compared to the unburn site that is 80% lichen (*Cladonia rangiferina*). Both burn sites had thaw depths >150cm by the end of the summer compared to the unburn site which was ~52cm. The burned site displayed similar pre fire stand density as the unburned sites and is within three kilometers of both sites, as such we do not believe the unburned site differed significantly from the burn sites pre fire.

3.3.2 Field data collection

For all our measurements we selected flat areas dominated by lichen (unburned site) or char (burned sites) surfaces, avoiding hummocks and hollows as these micro-topographical features are known to alter soil thermal regimes [*Wright et al.*, 2009]. These flat areas are representative of the studied sites where hummocks covered 5% (2000 burned site), 14% (2007 burned site) and 6% (unburned site) and hollows are mostly absent (burned sites) or covered 7% (unburned site) of the plateau area.

Soil temperature loggers (Pendant HoboProV2, Onset Corp.) were installed at a depth of 10cm and 40cm in two locations in each site during summer 2015. Air temperature was measured using a temperature logger at 1m above the ground, shielded from direct sunlight. Soil temperature time-series were obtained by recording measurements at two-hour intervals. Temperatures were averaged between the two loggers within each site due to periodic sensor failure, which occurred during the month of July in the 2000 burn site. Air and soil temperatures were measured from May 2016 to July 2017. Soil moisture sensors (Hobo Smart Sensor 10HS) were installed in July 2016 at 25 and 35 cm and recorded soil moisture every 20 minutes until July 2017.

Soil respiration fluxes were measured biweekly during the summer (Julian day 150 - 300) using static chambers. In each site, six collars (diameter = 0.4m) were installed 10cm into the soil. While our soil respiration measurements account for heterotrophic and autotrophic sources we consider the autotrophic component to be low due to the absence of vascular vegetation in our selected locations and the trenching of shallow roots during collar installation. Flux measurements started two weeks after collar installation using PVC chambers (volume = 37.7L) sealed with adhesive tape to the collars. A thermometer inside the chamber measured temperature in the headspace and carbon dioxide concentrations were recorded every 1.6 s for 2-3 minutes using a PP Systems EGM-4 portable infrared gas analyzer (IRGA; Amesbury, Massachusetts). The CO_2 flux rate was calculated as the slope of the linear relationship between headspace CO_2 concentrations and time. We excluded any slopes with r^2 less than 0.8, which represented less than 3% of our measurements. Soil temperature at 5, 10, and 40cm depth was recorded using a handheld thermometer during CO_2 flux measurements. We used soil temperatures from the loggers at 40cm when thermometers were not available (May). A comparison between thermometers and logger data at 10cm and 40cm showed no significant difference between both methods (t=0.9235, p<0.05). Soil moisture was also measured at the soil surface in four locations adjacent to the collar using a Delta-T HH2 portable soil moisture meter. Soil moisture was examined as a possible control on respiration. No relationship between soil moisture and soil respiration was found in either the burned or the unburned site (Figure

S2.1). Therefore, soil moisture was not a controlling factor and thus only soil temperature was used in soil respiration modelling. Several studies have shown both soil temperature and soil moisture to be an important control on soil respiration. However, there are controversial findings about the relative importance of soil moisture in controlling soil respiration in peatlands, with the relative importance of it in drier peatlands like peat plateaus being unknown [*Lafleur et al.*, 2005]. Although Natali et al. (2015) showed a positive correlation between monthly soil respiration and water table depth in a tundra upland, in peat plateaus water content at depth is relatively constant (Figure S3.1) and in general CO_2 production is most sensitive to moisture changes in the uppermost portions of the peat profile [*Lafleur et al.*, 2005]. In our study soil moisture content was relatively stable in the upper layers, thus the little change in heterotrophic respiration with soil moisture is not unlikely.

3.3.3 Modeling and statistical analysis

We examined the relationship between respiration and soil temperature using regression analysis. Soil moisture was not found to be a predictor of respiration (Figure S3.1) and was excluded from the analysis. Data from both burned sites was combined to evaluate the relation between temperature and soil respiration. Soil temperature and soil moisture did not differ between the two burned sites (*p>0.05,* Figure 3.2a and b, paired t-test), rendering confidence in combining data points from the two burned sites. Soil temperature from the six collars was averaged for each site in during each measurement day for the analysis. We used repeated measures ANOVAs to determine differences between soil respiration in burned and unburned sites.

We examined the temperature dependence of ecosystem respiration using nonlinear least square regression during the growing season using the following one- and two-depth soil temperature models:

(1) $R = (a * b^{(T5/10)} * BURN) + (c * b^{(T5/10)} * UNBURN)$

(2) $R = (a * b^{(T10/10)} * BURN) + (c * b^{(T10/10)} * UNBURN)$

(3) $R = (a * b^{(T40/10)} * BURN) + (c * b^{(T40/10)} * UNBURN)$

(4) $R = [(a * b^{(T5/10)} + c * b^{(T40/10)}) * BURN] + [(d * b^{(T5/10)} + e * b^{(T40/10)}) * UNBURN]$

(5) $R = [(a * b^{(T10/10)} + c * b^{(T40/10)}) * BURN] + [(d * Q_{10}^{(T10/10)} + e * b^{(T40/10)}) * UNBURN]$

where R is the rate ecosystem respiration (µmol C m^{-2} s^{-1}), a,c,d,e are fitted parameters from the regression representing the rates of ecosystem respiration at 0°C (µmol C m^{-2} s^{-1}) from the associated temperature depth, b is the Q_{10} - the temperature dependence of ecosystem respiration, T is the soil temperature at either 5, 10, or 40cm (°C) beneath the soil surface, and BURN/UNBURN are dummy variables of either 1 or 0 depending on the site the model is acting upon. The a, b, c, d, and e parameters were defined by using nonlinear lead squares (nls) package in R [*Bates and Chambers, 1992*].

The Akaike information criterion (AIC) was used to select the best model for both the burned and the unburned sites over the use of an R^2 due to its likelihood to exhibit an extreme bias to highly parameterized non-linear models [*Spiess and Neumeyer*, 2010]. The contribution of deep respiration during the growing season was calculated using the b * $Q_{10}^{(T40/10)}$ component of the model and dividing that by the total respiration (R). We compared the contribution of shallow (10cm) and deep (40cm) soil to soil respiration between sites the most parsimonious model and the soil profile temperatures measured near the collars during flux measurements.

Estimates of the cumulative soil respiration during the growing season at each site was calculated with the best fit model using soil temperature time-series from the loggers (Figure S3.2). Soil temperature was averaged from two loggers in the unburned site and from four loggers in the two burned sites. Model coefficients plus standard error was used for statistical bootstrapping with 300 iterations to determine an instantaneous flux rate for each 2-hour interval. These rates were then used to calculate cumulative respiration over the growing season for each site (JD 181 – 259).

3.4 Results

3.4.1 Soil Temperature

Burned sites had deeper active layers and warmer soils than the unburned site. During the summer period (Julian Day 150 – 250), soil temperature at 10 and 40cm was significantly higher in the burned sites than in the unburned site ($p<0.05$, paired t-test), although no differences between sites were observed nearer to the surface at 5 cm. Burned soils were over 2°C warmer, and over 4°C warmer at 10 and 40 cm respectively in the burned sites mid-summer (Figure 3.2a and Figure 3.2b).

Soil temperatures decreased with depth during the growing season and the difference between depths became smaller towards late fall, when the gradient with depth ended and temperatures at 10 and 40 cm converged and remained similar during the winter months. This occurred with a delay of about 25 days in the burned sites in comparison to the unburned site.

3.4.2 Soil respiration

Soil respiration had a seasonal response peaking towards the middle of the growing season in all sites. Rates were significantly higher in the unburned site than in both burned sites during the growing season ($F=11.42$, $p<0.05$, Figure 3.3a) and no significant differences occurred between the two burned sites ($F=0.025$, $p>0.05$, Figure 3.3a). Fluxes in the burned sites increased from (average ±1 standard deviation) 0.60±0.28 µmol C m^{-2} s^{-1} in early summer to 0.90±0.21 µmol C m^{-2} s^{-1}in late summer whereas these values were 0.88±0.48 and 1.1±0.28 µmol C m^{-2} s^{-1} in the unburned site, respectively. Overall, soil respiration was 27% greater on average in the unburned site than in the burned sites during the growing season period and the maximum difference occurred the early summer, at 46%. Soil temperature was exponentially related to soil respiration in both the burned and unburned site (Figure 3.3b) supporting the use of an exponential equation to model soil respiration.

3.4.3 Soil respiration models

The results of the regression models between respiration and temperature are presented in Table 1 and 2. The model that includes soil temperatures at both 10 and 40cm depth (model 1) was selected as the best fit based on its AIC value (Table 3.1 and 3.2). Incorporating the 40 cm soil temperature improved the model performance (compare model 1 with 2 and model 3 with 4) with the model containing only the 40 cm temperature performing the poorest (model 5). The models including the soil temperature at 10 cm fitted better to the data (model 1 and 2). The temperature dependence of soil respiration on temperature (Q_{10}) obtained from the modelled *b* parameter was 2.56±0.3. The most parsimonious model (model 1), which included, two temperatures was able to accurately model soil respiration during the growing season (Figure 3.4).

3.4.4 Deep peat contribution to CO_2 release

The burned sites had nearly 4 times as much respiration from depth (40cm) where soils thawed two weeks earlier than in the unburned site.

The proportion of soil respiration coming from deep layers in the burned sites increased 15% during the growing season from, ~36% in the early summer to ~55% in the late fall. This is, contrast to the unburned site that only experienced a 5% increase in respiration during the growing season, from ~9% in the early summer to ~14% in the late fall. (Figure 3.5).

We calculated the cumulative carbon loss from shallow and deep soil in the burned and unburned sites (Figure 3.6). While total CO_2 release was greater in the unburned site compared to the burned site during the growing season 83.92 ± 1.79 gC m^{-2} vs 76.58 ± 1.25 gC m^{-2} during the growing season (JD 150-258), deep peat sources (40cm) contributed substantially in the burned sites with a cumulative loss of 32.11 ± 0.97 gC m^{-2} compared to 8.17 ± 0.96 gC m^{-2s} in the unburned site (Figure 3.6b). By the end of the season this represents nearly 4 times as much carbon loss from deep soils to the atmosphere in burned sites.

3.5 Discussion

Recent research has shown that wildfire within the discontinuous permafrost zone can cause a deepening of the active layer in boreal permafrost peat plateaus [*Viereck et al.*, 2008; *Brown et al.*, 2015] which exposes large expanses of previously frozen soil organic carbon to microbial decomposition. Furthermore, not only do active layers deepen post-fire, deep (40cm) soil temperatures can rise as much as 4°C within permafrost peat plateaus [*Gibson et al. in prep*] which has been shown to stimulate soil respiration [*Davidson and Janssens*, 2006]. Therefore, in this study, our objective was to determine if fire induced deepening of the active layer leads to increased total soil respiration compared to unburned sites, and if it stimulates the respiration of old, previously frozen soil organic carbon. In order to do this, we explored the use of a two-temperature (shallow and deep temperature) soil respiration model, as a tool that may more accurately model soil respiration and allow us to partition respiration into both shallow and deep components.

Our first objective aimed to compare soil respiration from both a burned and unburned permafrost peat plateau. We hypothesized that, due to warmer temperatures and deeper active layers, the burned sites would have a higher rate of respiration compared to the unburned site. Surprisingly, the unburned site was found to have the higher rate of soil respiration at approximately 0.83±0.44 μmol C m^{-2} s^{-1}, compared to 0.66±0.30 μmol C m^{-2} s^{-1} in the unburned site, averaged over the course of the growing season (JD150-300). We found that the unburned site had consistently higher overall rates of respiration than the burned sites until mid-fall when rates of respiration were near equal and a third of what they were during peak growing season. We attribute higher soil respiration rates in the unburned site to the source and quality of carbon that is being respired in the burned versus the unburned site. During a fire event, the top 5-20cm of peat is combusted [*Turetsky et al.*, 2011]. This peat has recent inputs of fresh, labile carbon that exhibits higher turnover rates, as opposed to carbon at depth that is more recalcitrant [*Bosatta and Ågren*, 1999; *Knorr et al.*, 2005]. Therefore, even a low severity fire that removes just 5 cm of organic soil is likely to result in the loss of the labile carbon pool. We attribute the observed lower respiration rates of the burned sites to soil microbes that favour these easily decomposable organic compounds found in the top layers. This results in the high rates of decomposition near the surface, and decreasing rates with depths as carbon pools become less labile [*Bergeron*, 2000]. Interestingly, despite our findings that post fire increases in soil temperatures do increase respiration depth, we conclude that this is not enough to overcome the much higher respiration rates of labile surface peat. As a result of this, we conclude that wildfire does not cause these ecosystems to become larger sources of carbon to the atmosphere.

Our results are contrary to that of Dorrepaal et al. (2009), who showed long-lasting sensitivities to climate warming by soil carbon towards the bottom of the active layer in subarctic peatlands. They found the fresh surface litter and plant-related respiration contributed proportionally less to total respiration than in the most other ecosystems because of the large, organic soil pool at depth. Our results also support the conclusion of increased respiration at depth due to increased soil temperatures though they do not support the notion the fresh surface litter inputs are not as significant in total respiration (Figure 6). Despite burned sites having a large proportion of total respiration coming from depth, the higher yearly respiration from the unburned site that has over 100cm less thawed organic soils highlights the importance of these fresh surface litter inputs.

Despite our results being limited to the growing season, we observed a trend towards increasing respiration from the burned sites towards the end of the season compared to the unburned site that displayed as faster rate in decline (Figure 3). During this time period (JD>280) we observed convergence of the surface temperature, but the temperatures at depth remained warmer in the burned site. We suspect that as a consequence of this, respiration remains higher at depth in the burned site during a period of time where surface respiration is relatively minimal. Although outside the scope of this study, if this trend were to continue during the winter months it could have a noticeable effect on the net carbon loss from these soils on an annual basis. As well, during this time period there is decreased fresh litter inputs as vegetation begins to senesce, likely contributing to decreased respiration at the surface in unburned sites.

Our second objective in this study was to examine models for soil respiration that used one or two temperatures. By using a two-temperature model we were able to effectively partition respiration into shallow (10cm) and deep (40cm) respiration. Most attempts at partitioning respiration have come from incubation studies (see [*Schädel et al.*, 2016]) or Δ14C measurements (see [*Dutta et al.*, 2006; *Hilasvuori et al.*, 2013]). To the best of our knowledge, our study is one of the first to model soil respiration with two temperatures. In general, a single temperature at a shallow depth is used to model soil respiration [e.g. *Chivers et al.*, 2009]. This is often because single temperature models that use only deep temperatures poorly represent the data. This was also observed in our study with the single temperature at 40cm being our worst performing model based on an AIC evaluation (Table 1). However, we show that when used in conjunction with a shallow temperature not only does a two temperature model effectively model the data (Figure 5) it can be used as a new way to model respiration from shallow and deep portions of the soil profile.

Our modeled temperature sensitivity of decomposition (Q_{10}) was similar to those seen in other studies that incubated active layer soils, 1.8 – 2.5. Here we applied the same Q_{10} value for both our shallow and deep soil contributions, though recognize that these two depths may have different sensitives as deeper, more recalcitrant carbon generally has a higher temperature sensitivity [*Dorrepaal et al.*, 2009]. However, due to the constraints of our models, it was necessary to set them as equivalent. Other studies that have examined temperature sensitives on peat plateaus have reported significantly higher Q_{10} values (example, [*Startsev et al.*, 2016], which reported 5.6 – 11.5) though this is not generally accepted as biologically valid, and therefore may be an artifact of data sampling and analysis. There are several environmental constraints that can affect the temperature sensitivity of organic material to decomposition.

While freezing severely limits microbial activity, upon thaw both chemical and biological factors can limit the temperature sensitivity of decomposition. Many incubations studies have shown a sharp decline in soil respiration rates within the first 3 weeks of the incubation, and an increase in the temperature sensitivity and respiration rates due to the depletion of the labile carbon pool [*Conant et al.*, 2008; *Xu et al.*, 2010]. This idea supports our findings that temperature sensitivities and respiration in a burned site would be lower due to the loss of the labile carbon pool during the fire event.

Our last objective was to derive estimates of deep respiration to quantify the release of thawed carbon in burned sites and assess the effects of disturbance on the carbon cycle in permafrost peatlands. Our finding that wildfire increased respiration from depth is consistent with the work of [*Dorrepaal et al.*, 2009] that showed that at least 69% of an observed increase in respiration with warming originated from carbon in deeper peat layers. In our study we show that up to 55% originates from deeper peat layers following a fire event, likely due to upwards of 100cm deeper active layers in the burned than unburned site, suggesting the mobilization of old carbon from depth. Collectively, warming of soils at depth, whether via climate warming or wildfire, is accelerating the respiration of subsurface carbon reservoirs to a much larger extent than previously thought [*Christensen et al.*, 1999; *Hogg et al.*, 1999].

Surprisingly though, we found that the unburned site had consistently higher overall rates of respiration than the burned sites. As discussed earlier, we attribute this to the source and quality of carbon that is being respired. As a result of this labile carbon pool we conclude that the increased respiration from depth isn't enough to overcome the importance of this labile pool. Regardless, wildfire does result in old carbon being incorporated into the current carbon cycle and given present day climate warming and permafrost thaw rates it is unlikely that this newly incorporate pool of carbon will be fixed and preserved in permafrost once again, representing a long term source of carbon in our atmosphere.

3.6 Conclusion

As fire frequencies and severity changes in northern boreal forests as a result of climate change, widespread permafrost thaw is likely to occur, and cause drastic changes to carbon cycling in these environments [*Schuur et al.*, 2015]. Our study shoes that when permafrost peat plateaus undergo fire, and subsequent active layer deepening [*Gibson et al. in prep*], they do not become larger sources of carbon to the atmosphere in the years to decades following the fire event due to increased soil respiration. This is despite having over 100cm more thawed soils and 4ºC warmer soils at depth. Our study disputes that contribution of soil respiration, from exposed, deep active layers, will be a greater source of carbon to the atmosphere than fresh surface litter.

This study revealed that despite higher temperature sensitivities at depth, burned permafrost peat plateaus actually had lower soil respiration than unburned peat plateaus. Furthermore, this occurred despite significantly higher proportion of respiration from depth in the burned site, highlighting the likely importance of fresh surface litter inputs that are burned off during the fire. These findings have important implications for prediction of future climate changes in permafrost peatland ecosystems. It is important to reconcile the discrepancy between modeled and observed increases in soil respiration because models that account for increased temperature sensitivities at depth, but neglect the loss of labile surface carbon during burning, will likely overestimate post fire soil respiration in permafrost peat plateaus. Ultimately, wildfire is leading to the loss of old soil carbon from depth that was previously stored in the terrestrial environment. As the extent of wildfire increases across the discontinuous permafrost zone, the rate of old carbon mobilization to the atmosphere will increase. What our results also highlight is the potential large soil carbon loss that could exists in the absence of fire, but in the presence of climate warming. If permafrost peat plateaus experience deepening of the seasonal active layer but also maintain the fresh sources of labile carbon inputs larger than expected rates of soil respiration could occur as carbon is mobilized from depth.

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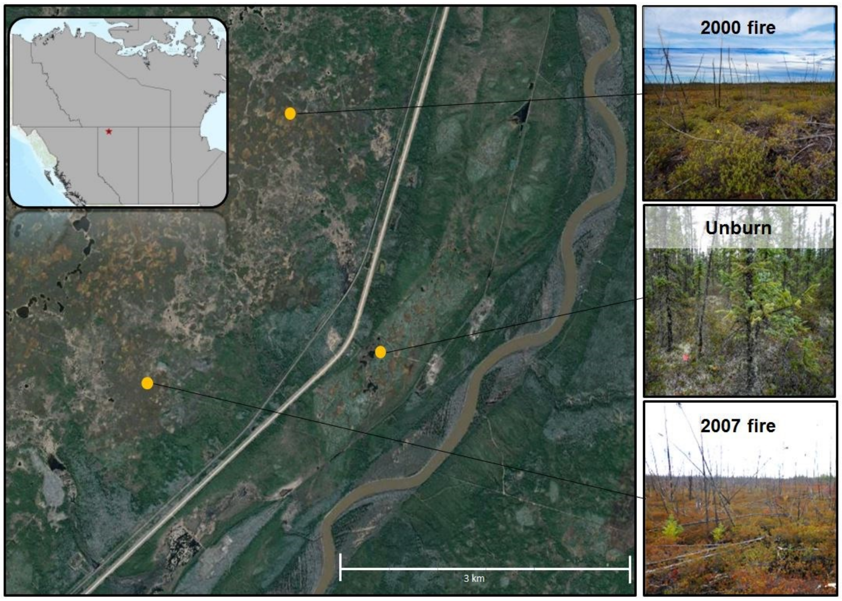

Figure 3.1: Peatland complexes and field sited in the Taiga Plains ecozone (inset) in western boreal Canada. Peatland plateaus within the Taiga Plains are a mosaic of treed permafrost peatlands and open, permafrost free thermokarst bogs. Study sites included two burned permafrost peat plateaus that burned during the summer of 2000 and 2007, and an unburn site

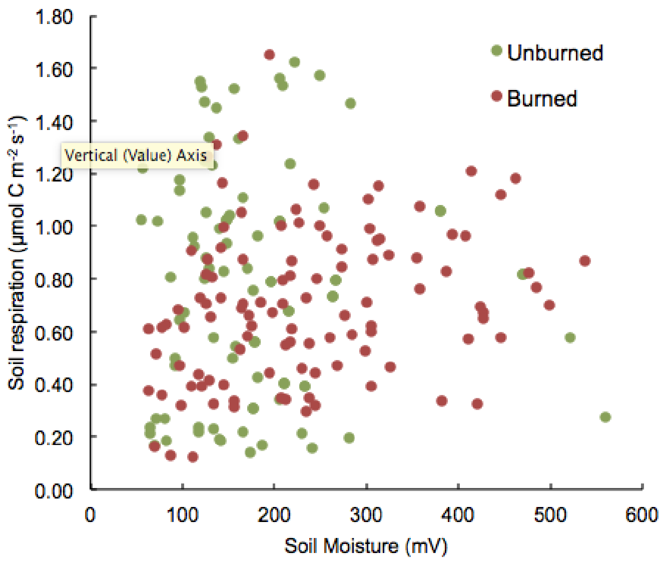

Figure S3.1: Soil moisture using a handheld soil moisture sensor and soil respiration. No relationship was found between soil respiration and moisture therefore soil moisture was not included in our models of soil temperature as a controlling factor.

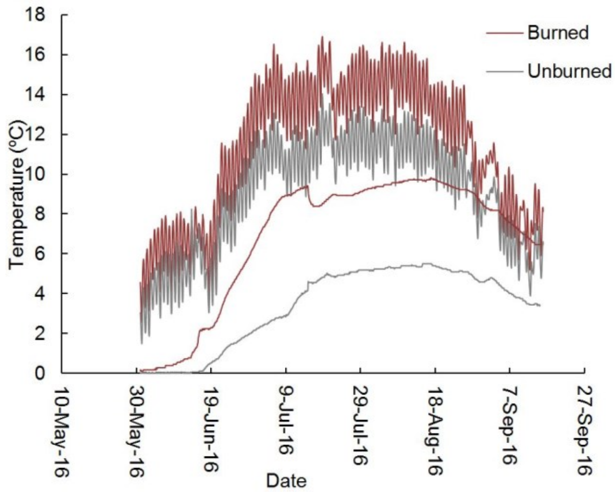

Figure S3.2: Soil temperature, in two hour intervals, from hobo temperature loggers from shallow, 10cm (upper lines), and deep, 40cm (lower lines), during growing season. Temperature is averaged from the two loggers located within each site.

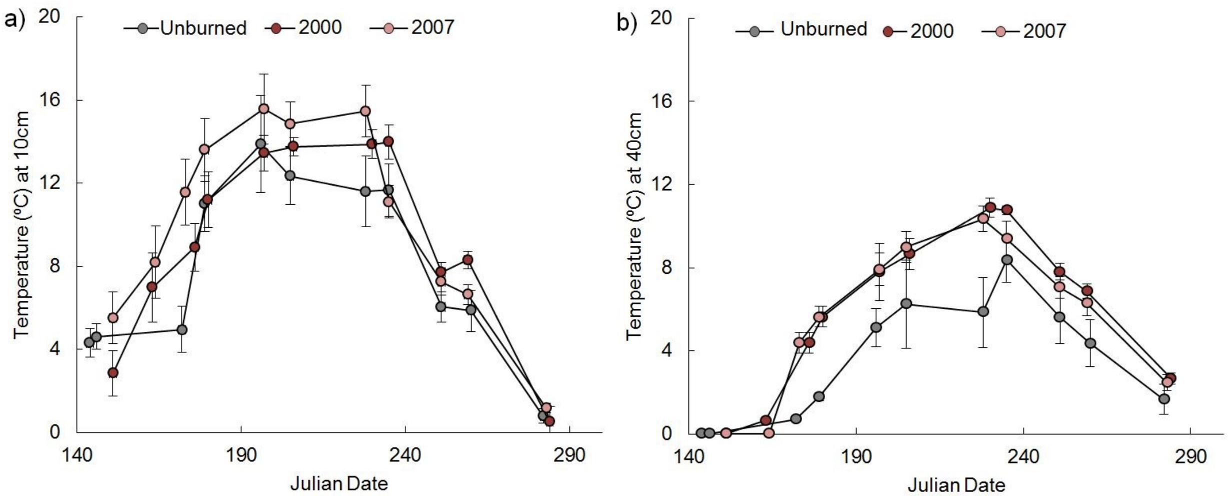

Figure 3.2: Soil temperature during growing season (JD 150-250). A) Soil temperatures at 10cm were significantly warmer in the burned than in the unburned (*t= 2.705, p<0.05*). B) Soil temperatures at 40cm were also significantly higher in the burned than in the unburned (*t=5.414,p<0.05*) during the 2016 growing season.

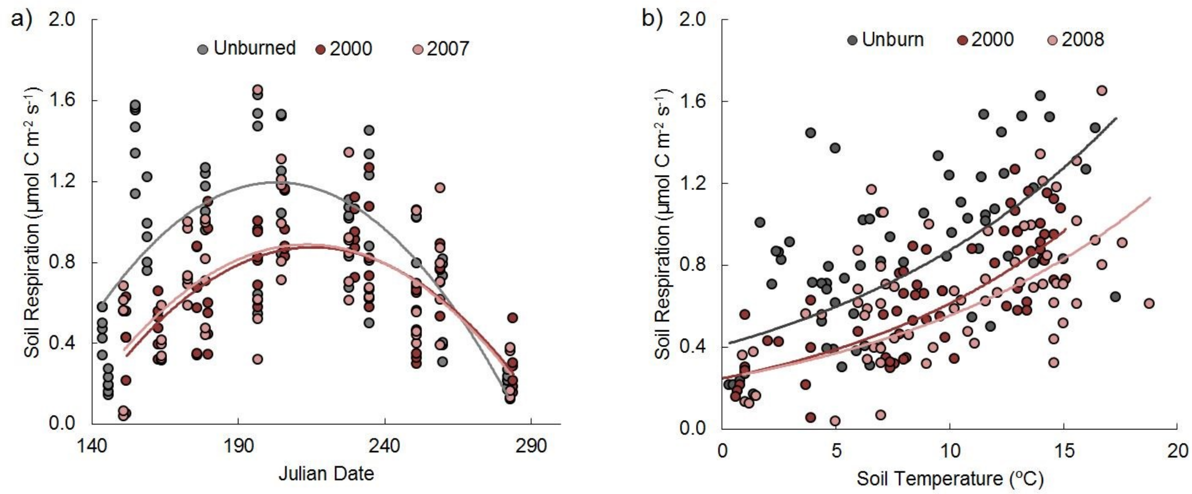

Figure 3.3: Raw soil respiration during the growing season (JD 148-300). A) Soil respiration during the season with a second order polynomic trend line. Unburned sites displayed higher respiration during the early summer while burned sites showed increased respiration late summer. B) Exponential relationship between soil temperature and soil respiration.

Table 3.1: Modeled equations for burned and unburned sites, arranged in order of decreasing model fit based on Akaike information criterion.

Model	*R2*	*AIC*
$R = [(a * b^{(T10/10)} + c * b^{(T40/10)})*BURN] + [(d * b^{(T10/10)} + e * b^{(T40/10)})*UNBURN]$	0.90	-34.21
$R = (a * b^{(T10/10)}*BURN) + (c * b^{(T10/10)}*UNBURN)$	0.88	-33.61
$R = [(a * Q_{10}^{(T5/10)} + c * b^{(T40/10)})*BURN] + [(d * b^{(T5/10)} + e * b^{(T40/10)})*UNBURN]$	0.88	-28.19
$R = (a * b^{(T5/10)}*BURN) + (c* b^{(T5/10)}*UNBURN)$	0.79	-16.40
$R = (a * b^{(T40/10)}*BURN) + (c * b^{(T40/10)}*UNBURN)$	0.70	-5.65

Table 3.2: Coefficients (±1SE) of variables in the best model based on an AIC evaluation.

Variable	**Coefficient**
a – Respiration at 0°C (10cm burned)	0.14 ± 0.05 µmol C m^{-2} s^{-1}
b – Q_{10}	2.56 ± 0.31
c – Respiration at 0°C (40cm burned)	0.14 ± 0.06 µmol C m^{-2} s^{-1}
d – Respiration at 0°C (10cm unburned)	0.30 ± 0.08 µmol C m^{-2} s^{-1}
e – Respiration at 0°C (40cm unburned)	0.05 ± 0.09 µmol C m^{-2} s^{-1}

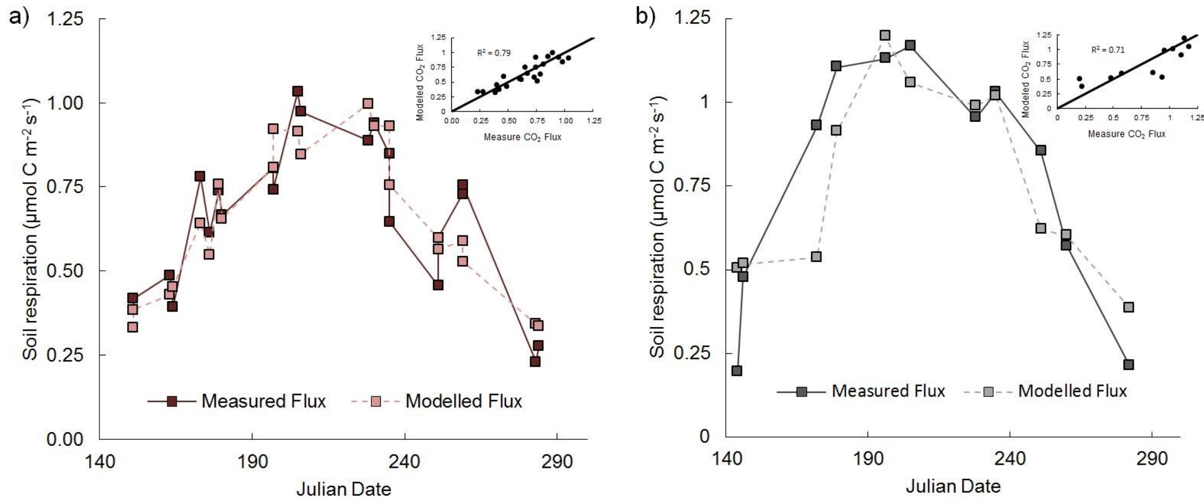

Figure 3.4: Modeled and measured soil respiration during the growing season (JD 140-300) in the burned (a) and unburned (b) sites. Model accurately predicted fluxes in the burned (R^2=0.79) and the unburned (R^2=0.71). Model performed most poorly at low temperatures and low fluxes, likely due to low n size at these conditions.

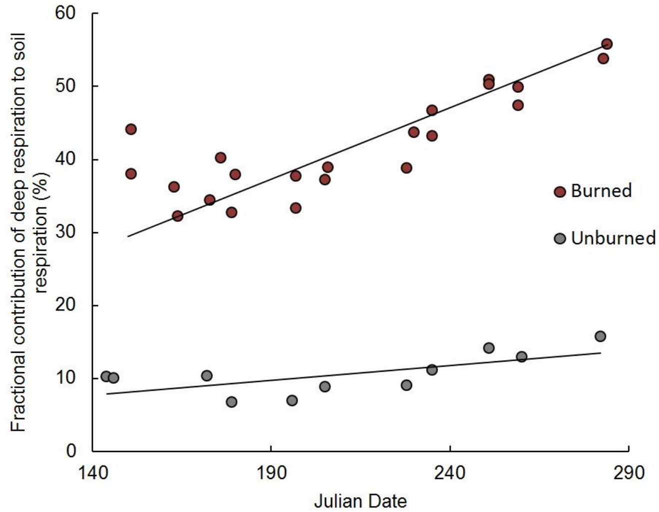

Figure 3.5: Fractional contribution of deep respiration to total soil respiration from the burned and unburned sites. Burned sites displayed increased respiration from depth later into the summer season whereas the unburned site remained relatively consistent, with only a minor increase.

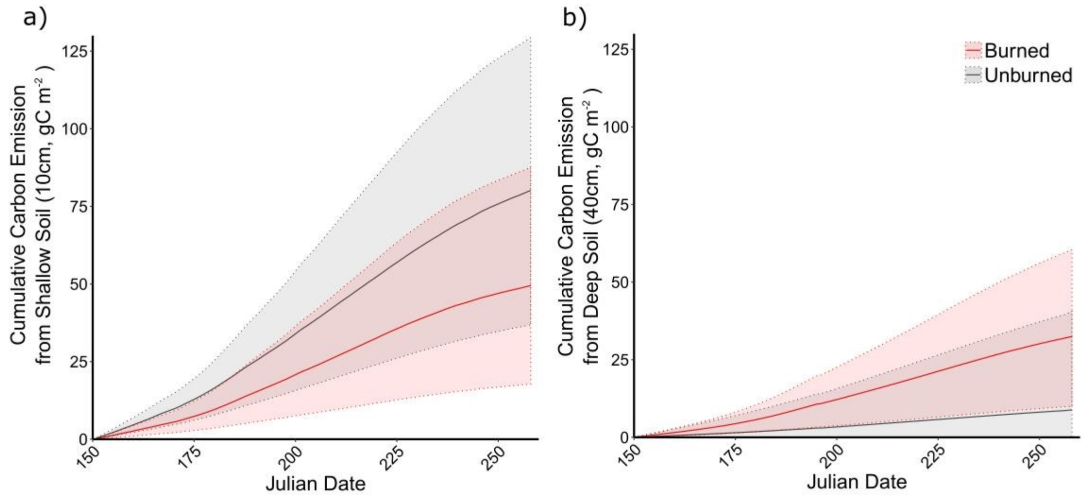

Figure 3.6: A) Cumulative carbon loss (gC m^2) from shallow soils (10cm) during the growing season (JD 150-258) with upper and lower bounds based on a Q_{10} of 2.56 ± 0.31. B) Cumulative carbon loss from deep soils (40cm) during the same period.

Chapter 4: Conclusion

The goal of this thesis was to investigate how wildfire impacts permafrost stability and carbon cycling in northern permafrost peatlands. This work was motivated by projected changes to the fire regime [*Amiro et al.*, 2009] and permafrost stability [*Viereck et al.*, 2008; *Baltzer et al.*, 2014; *Mamet et al.*, 2017] within the discontinuous permafrost zone. Furthermore, the vast majority of studies that have begun to explore this issue have concentrated in Alaska peatlands, where peat properties and depths differ, thus making the extrapolation of their results to the circumpolar boreal region difficult. Therefore, this research focused on the discontinuous permafrost zone of the Taiga Plains in northwestern Canada where permafrost is found principally within peatlands where permafrost underlies peat plateaus that rise 0.5-1 m above surrounding permafrost free wetlands. These ecosystems also contain some of the most carbon rich soils within the circumpolar north [*Hugelius et al.*, 2013], thus making their response to disturbance very important for future climate conditions.

The first objective of this research was to understand over what temporal scale wildfire affects active layer depths. I found that wildfire's effect on active layer depths was twofold. On the short term, 0-10 years post fire, it causes a continuous deepening of the active layer. Follow this time period, as vegetation begins to recovers, the rate of thaw begins to decrease and active layers recover. However, a longer term effect of fire was also the formation of taliks, which were prominent from 10-20 years post fire. Therefore the net effect of fire can be seen for ~30-40 years at which time active layer depths have returned to pre-fire conditions and the proportion of taliks within the site is similar to that a unburned peat plateaus. This has important implications for the stability of the permafrost core and its net energy balance. We propose that due to the presence of these taliks, sufficient energy isn't able to escape during the winter which causes thawing at the edge of the permafrost core (objective 2).

The second object was to understand the extent that wildfire accelerates thermokarst development. I found that perturbed soil thermal conditions due to active layer deepening are likely responsible for a near tripling in recent thermokarst area within burned areas. I found that despite wildfire only affecting ~25% of permafrost peatlands within the study area, it is responsible for 25% of all thaw or 2,700km^2 of thawed permafrost thaw. Consequently, we conclude that it is important to incorporate the future role of wildfire in permafrost stability in models for circumpolar boreal regions. Also, this thaw has important implications for traditional land-use, regional hydrology and water quality, and soil greenhouse gas emissions (objective 3).

The third object was to understand how soil respiration is affected following wildfire. I found that as a result of wildfire removing the top 5-20 cm of organic soils during a fire event [*Turetsky et al.*, 2011] removes the labile carbon pool burn sites do not become greater soils of carbon to the atmosphere. This is despite warmer soil temperatures at depth and increased active layers (ie. larger thawed soil carbon pools). However, by being one of the first to use a two temperature soil respiration model that incorporated a shallow and deep temperature we were able to partition the respiration and understand the magnitude of shallow and deep respiration (objective 3).

My final objective was to understand the relative proportion of respiration from depth in a burned site compared to an unburned site as this represents the respiration of deep, old, previously frozen carbon stores. I found that due to higher temperatures at depth deep respiration represented over 50% of all respiration within a burned site compared to just 14% of all respiration, representing over a fourfold increase in respiration from depth as a result of wildfire. This culminates in ~31.87 ± 13.45 kg C m^{-2} loss during the growing season (JD 150-258) from a burned peat plateau and only 8.60 ± 12.13 kg C m^{-2} from an unburned peat plateau. This highlights an importation potential phenomenon, that if active layers and soil temperatures increase with climate warming, in the absence of fire, an overall increased respiration could occur due to a potential 4 fold increase in respiration from depth.

Taken collectively, these conclusions show that wildfire can act as a major driver of change for both permafrost stability and carbon cycling within the discontinuous permafrost zone. Globally, permafrost peatlands cover 8% of the circumpolar world and store ~13Pg of soil carbon. By understanding their response to wildfire, we can better predict future permafrost conditions and carbon losses to the atmosphere. This will help us to make informed management and development decisions in the future.

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