

Impacts of forest fire on permafrost in the discontinuous zones of northwestern Canada

A thesis submitted in partial fulfillment of the requirements for the
Doctorate in Philosophy degree in Geography

Jean Holloway

Department of Geography, Environment and Geomatics
Faculty of Arts
University of Ottawa

Supervisor:
Dr. Antoni Lewkowicz, University of Ottawa

Examining committee:
Dr. Denis Lacelle, University of Ottawa
Dr. Sharon Smith, Geological Survey of Canada and University of Ottawa
Dr. Chris Burn, Carleton University

Dr. Fritz Nelson, Michigan State University

© Jean Holloway, Ottawa, Canada, 2020

ABSTRACT

Climate change is causing increases in the frequency, severity, and extent of fires in the boreal forest, which in turn is expected to change historical cycles of permafrost response and recovery to disturbance. A review of recent literature (forming part of this thesis) shows that there are disparities in current knowledge of post-fire permafrost response. First, the majority of studies on permafrost-fire interactions have been conducted in Alaska, leaving regional gaps for the boreal forest across Canada. Second, there are limited direct measurements of certain variables which affect post-fire permafrost dynamics. These include snow depth, burn severity, and soil moisture, as well as ground ice content and quantified subsidence and thermokarst development. Third, the majority of post-fire permafrost studies address near-surface impacts, neglecting permafrost conditions at depth. Finally, there is a lack of long-term information and regional investigations over a broad range of environmental conditions, particularly how permafrost responds across a variety of ground ice contents. This thesis addresses these knowledge gaps through in-situ measurements and analysis of permafrost conditions following fires occurring over the last half century and in particular in 2014 and 2015, along a 650 km latitudinal transect spanning the discontinuous zones, from isolated patches (57.8°N) to extensive discontinuous permafrost (63.1°N), in northwest Canada.

A variety of monitoring techniques were used to evaluate permafrost change, including ground and air temperature measurements, direct current electrical resistivity tomography (ERT) surveys, measurements of frost table depth, snow depth, organic layer thickness, burn severity, and ground subsidence. Samples of frozen and unfrozen soil were collected by coring or pit digging, and laboratory analyses conducted to establish soil characteristics. Laboratory experiments were also performed to establish a relationship between resistivity and temperature, and to generate a threshold between frozen and unfrozen soil that could assist in the interpretation of ERT surveys.

A total of 68 sites along the Mackenzie Highway in northern Alberta and southern Northwest Territories (NWT) were examined to evaluate permafrost change due to climate warming and forest fire since an initial survey in 1962. The transect extends through the isolated patches and sporadic discontinuous permafrost zones, including 11 sites which burned at various times between 1971 and 2012. Overall, there has been significant permafrost degradation, especially at sites with thin organic layers and coarse-grained soils. This occurred preferentially

at the southern end of the transect, where nearly 2°C of climate warming has occurred, such that even undisturbed sites experienced degradation. However, permafrost has persisted at about half of the sites where black spruce (*Picea mariana*) canopies with organic layers generally >40 cm thick overlie fine-grained sediments. Permafrost even persisted at the majority of burned sites, but greater frost table depths were observed at those which were burned in 2012.

A second transect was established to examine permafrost change following the abnormally severe fire year of 2014 (and more limited fires in 2015) that affected sites across a wide range of conditions in the southern NWT. Eleven monitoring sites were established in the sporadic and extensive discontinuous zones between 2015 and 2016, and annual field surveys were conducted through to 2019, including the first repeat ERT surveys conducted following fire. Permafrost change occurred at all sites, including unburned ones, indicating the ongoing impacts of climate change in the region. Snow-depth days, maximum snow depth, and the nival offset were all greater at burned sites. Permafrost change was more pronounced at burned sites, with greater relative decreases in average apparent resistivity and increases in frost table depths and ground temperatures, particularly at sites with low gravimetric moisture content, coarse soil textures, and organic layers <40 cm thick. These changes are pronounced in the near-surface (<5 m depth), with deeper permafrost appearing relatively unaffected within the 5-year post-fire time-frame.

The field observations indicate that permafrost can still persist following fire at a significant percentage of locations in the discontinuous zone. However, slow degradation is occurring at both burned and unburned sites due to the warming climate, and particularly at dry sites with coarse-grained soils and thin organic layers. Post-fire permafrost change is evident at sites which burned in the last 10 years, but over the long-term, frozen ground appears resilient to fire, with characteristics like active layer thickness returning to pre-fire levels. Similarly, cold permafrost on the taiga shield is resilient to fire, even with thin residual organic layers. At high ice-content sites, however, where ground subsidence and thermokarst develop, water inundation and permafrost thaw can occur, particularly in areas which have been severely burned. This thesis underlines, therefore, the importance of monitoring and modelling a variety of landscape types to establish post-fire permafrost impacts and temperature trajectory, and more specifically the effects of heterogeneity of drainage conditions, substrate, and organic layer thicknesses on the fate of permafrost in the boreal forest.

ACKNOWLEDGEMENTS

This thesis would not have been possible without the support I've received both in terms of personal encouragement and backing for the project as whole. This is a culmination of a lot of hard work and assistance from numerous parties, contributions which are invaluable. This journey has had a lot of ups and downs for me, and it truly wouldn't have been possible without all the support and guidance I've been graced with throughout the process.

First and foremost, I would like to thank the people of the Dehcho and North Slave regions who granted me access to work on their traditional land. Specifically, I would like to thank the communities of Fort Providence, Kakisa, Behchoko, Hay River, and Yellowknife for welcoming me and working with me throughout the course of my project. Mahsi Cho!

The most gratitude goes out to my supervisor, Dr. Antoni Lewkowicz, for making this entire project possible. You've imparted a wealth of knowledge to me, and particularly taught me to be a better writer and think critically. *Nonsense can never be right*. I will be forever grateful for your mentoring and encouragement over the years, particularly when times were tough. You've given me the wonderful gift of travel all over the world, but particularly to Yellowknife and the boreal forest at large, which I've come to love so much (enough to get a tattoo in fact). I hope to one day possess the same enthusiasm you have for arctic research, for teaching, and for life in general. I would also like to thank the other investigators of the CIMP project, without who this project wouldn't have come to realization. In particular, Dr. Merritt Turetsky and Dr. Jennifer Baltzer who provided logistical support, guidance, and mentorship throughout my thesis. Many thanks go to all coauthors and collaborators on manuscripts found within this thesis, but especially Dr. Tom Douglas for teaching me how to permafrost core. A special thank you to Dr. Stephen Wolfe and colleagues at the Geological Survey of Canada for providing in-kind support to this project, supplying data for Lucky Lake sites, and helping me frost probe. Next, to all the members of the fire crew and those who assisted me in the field over the years – I couldn't have done it without you! I would also like to thank my lab mates and other graduate students in the Department for mentoring, encouragement, and friendship over the course of my degree.

A very special acknowledgement goes out to my family and friends, specifically for putting up with me talking about nothing but permafrost, and for lifting me up when I needed it. To my mother, Jamie, thank you for the support and the motivation, and for reading my papers even if you didn't understand a word I was saying. To my sister, Tracy, you have been my rock

and best friend these last five years, and always. Thank you for singlehandedly getting me through my comprehensive exams! And finally, to my father, Arthur. Thank you for your unwavering support. After countless conversations, you are still as excited as ever about what I am studying and where I am headed in my life, academic or otherwise.

I would like to thank my committee members, Dr. Christopher Burn, Dr. Denis Lacelle, and Dr. Sharon Smith for their advice and guidance throughout my PhD program. In particular, to Denis for allowing me to use his environmental test chamber to conduct my permafrost experiment, to Chris for inviting me to submit a paper for the Transactions of the IPA special issue, and to Sharon for all the helpful chats. I would like to thank Dr. Fritz Nelson for serving as the external examiner for my thesis, and for his feedback that was extremely useful in improving my manuscripts.

This research would not have been possible without financial and logistical support from the Government of the Northwest Territories (NWT), the NWT Cumulative Impact Monitoring Program, the Natural Sciences and Engineering Research Council, the Northern Scientific Training Program, the W. Garfield W. Weston Foundation for Northern Research, the Royal Canadian Geographic Society, the Ontario Graduate Scholarship, and the University of Ottawa.

TABLE OF CONTENTS

ABSTRACT	ii
ACKNOWLEDGEMENTS	iv
TABLE OF CONTENTS	vi
LIST OF TABLES	ix
LIST OF FIGURES	x
CHAPTER 1: INTRODUCTION.....	1
1.1 Introduction	1
1.2 Research objectives and thesis organization.....	4
1.3 Authorship and co-authorship contributions	11
1.4 References	12
CHAPTER 2: IMPACT OF WILDFIRE ON PERMAFROST LANDSCAPES: A REVIEW OF RECENT ADVANCES AND FUTURE PROSPECTS	20
2.1 Introduction	20
2.2 Post-fire changes to active layer thickness and the near-surface thermal regime	23
2.3 Post-fire thermokarst development in ice-rich permafrost.....	26
2.4 Post-fire permafrost recovery and resilience.....	29
2.5 Hydrological and biogeochemical impacts of wildfire in permafrost landscapes	35
2.6 Discussion: Priorities for future research	38
2.7 Conclusions	41
2.8 References	42
CHAPTER 3: HALF A CENTURY OF DISCONTINUOUS PERMAFROST PERSISTENCE AND DEGRADATION IN WESTERN CANADA.....	53
3.1 Introduction	53
3.2 Study area	54
3.3 Methods	57
3.4 Results	59
3.4.1 Climate changes.....	59
3.4.2 Changes in near-surface permafrost distribution.....	62
3.4.3 Frost table depths.....	64
3.4.4 Geophysical surveys.....	65
3.4.5 Forest fire.....	67
3.5 Discussion	68
3.5.1 Loss and persistence of permafrost	68
3.5.2 Permafrost and forest fire	72
3.6 Conclusion.....	74
3.7 References	75

CHAPTER 4: FIELD AND LABORATORY INVESTIGATION OF ELECTRICAL RESISTIVITY-TEMPERATURE RELATIONSHIPS, SOUTHERN NORTHWEST TERRITORIES.....	81
4.1 Introduction	81
4.2 Field sites.....	82
4.3 Methods	84
4.3.1 Field measurements	84
4.3.2 Permafrost sampling and laboratory measurements	85
4.3.3 Data processing.....	85
4.4 Results and Discussion	86
4.4.1 Variation in resistivity with temperature	86
4.4.2 ERT tomograms and temperature contour maps	87
4.4.3 Resistivity and temperature measurements in the field and laboratory	89
4.4.4 Comparison of ERT tomograms and temperature contour plots	91
4.5 Conclusions	92
4.6 References	93
CHAPTER 5: CHANGES TO PERMAFROST IN THE 5 YEARS AFTER FOREST FIRE, SOUTHERN NORTHWEST TERRITORIES, CANADA	96
5.1 Abstract.....	96
5.2 Introduction	96
5.3 Study Area	98
5.4 Methods.....	101
5.4.1 Site Selection	101
5.4.2 Temperature and snow depth measurements.....	101
5.4.3 Direct current electrical resistivity tomography surveys.....	105
5.4.4 Frost table depths.....	106
5.4.5 Ground subsidence	106
5.4.6 Burn severity measurements.....	107
5.4.7 Soil sampling and laboratory measurements	108
5.5 Results	109
5.5.1 Post-fire changes in air and near-surface temperatures	110
5.5.2 Post-fire changes in frost table depths.....	112
5.5.3 Post-fire changes in resistivity.....	116
5.5.4 Impact of burn severity on permafrost	121
5.5.5 Contribution of ground ice to post-fire permafrost change	124
5.5.6 Post-fire changes in snow and impacts on permafrost	128
5.6 Discussion	130
5.7 Conclusion.....	138
5.8 References	139
CHAPTER 6: CONCLUSIONS	146
6.1 Summary and conclusions	146
6.2 Key contributions	148

6.3 Limitations and future research directions	149
6.4 References	153
THESIS BIBLIOGRAPHY	156
APPENDICES	177
APPENDIX A: SUPPLEMENTAL INFORMATION FOR CHAPTER 2	177
APPENDIX B: SUPPLEMENTAL INFORMATION FOR CHAPTER 3	186
APPENDIX C: SUPPLEMENTAL INFORMATION FOR CHAPTER 5	195

LIST OF TABLES

Table 2-1: Factors that control post-fire permafrost recovery	30
Table 3-1: Linear trends in climate variables over various periods for Hay River, NT and High Level, AB	61
Table 3-2: Presence of near-surface permafrost in the study area	63
Table 3-3: Information about permafrost conditions and fire history at sites which burned between 1962 and 2017	68
Table 4-1: Literature resistivity values for the boundary between frozen and unfrozen soils	82
Table 4-2: Field site information	84
Table 4-3: Relationship between resistivity and temperature	86
Table 5-1: Instrumentation at each monitoring site	103
Table 5-2: Descriptive index of burn severity	108
Table 5-3: The difference in ground surface elevation between ZipLevel surveys taken in September, 2016 and September, 2018	125
Table 5-4: Summary table of the conditions at each site between 2015-2019	131

LIST OF FIGURES

Figure 1-1: Map of the study region	7
Figure 2-1: Locations of recent studies on the impacts of wildfire on permafrost landscapes ...	22
Figure 2-2: Observations of post-fire active layer thickness from recent literature for boreal and tundra sites	25
Figure 2-3: Photos of post-fire thermokarst development in areas where permafrost has high ice content	27
Figure 2-4: Conceptual diagram showing the potential post-fire trajectories for permafrost without excess ice and hence limited or no surface subsidence during thaw	34
Figure 2-5: Conceptual diagram showing the potential post-fire trajectories for permafrost containing excess ice and subject to surface subsidence during thaw	35
Figure 3-1: Study transect along the Mackenzie Highway showing changes in permafrost conditions between 1962 and 2017-2018 and permafrost zonation	55
Figure 3-2: Mean annual air temperature at Hay River, NT from 1945 to 2017 and High Level, AB from 1967 to 2017	56
Figure 3-3: Proportion of the 68 sites in each change category	63
Figure 3-4: Relation between permafrost change (persistence or loss) from 1962 to 2017-2018 and (a) organic layer thickness and (b) substrate	64
Figure 3-5: Frost table depths: (a) Comparison of site values for 1962 and 2017-2018; (b) Intra-site ranges of adjusted frost table depths for 2017-2018	65
Figure 3-6: 160 m long permafrost survey undertaken at site 54 (59.9°N) on August 27, 2017: (a) Frost table depths measured at 2-4 m intervals (blank areas indicate no frost table within 110 cm of the surface), and; (b) ERT survey shown as model blocks	66
Figure 3-7: The study region and fire scar polygons from 1931-2017 for Alberta and 1965-2016 for NWT	67
Figure 4-1: Study transect along the Mackenzie Highway showing site locations and soil type, permafrost zonation, and ecozones	83
Figure 4-2: Relationship between resistivity and temperature for measurements in the field and laboratory	88

Figure 4-3: 160 m long ERT surveys and frost table depths for (a) site UB-1, (b) site SS33-9a and (c) site ZF20-6b-LK	90
Figure 4-4: Two-dimensional contour plots of predicted temperatures at UB-1 (a), SS33-9a (b), and ZF20-6b-LK (c)	91
Figure 5-1: Map showing the study area in the southern Northwest Territories	99
Figure 5-2: Mean annual air temperature (MAAT) at Hay River, NT (circles) and Yellowknife, NT (diamonds) from 1989 to 2019	100
Figure 5-3: Boxplots of mean annual air temperature (MAAT), mean annual ground surface temperature (MAGST), and mean annual ground temperature at 1 m depth (MAGT) for all burned and unburned sites for 2016-2019	110
Figure 5-4: Changes in mean annual ground temperature (MAGT) at 1 m depth at the monitoring sites in relation to: (a) average grain size in the soil column, (b) average organic content, and (c) average gravimetric moisture content of organic soils in the active layer	113
Figure 5-5: Ground surface thawing degree days (TDD _s) and freezing degree days (FDD _s) for all burned and unburned sites between 2016 and 2019	114
Figure 5-6: Average first quartile (Q1) frost table depths for each site and years based on: (a) the average grain size in the soil column, (b) the average organic content, and (c) the average gravimetric moisture content of organic soils in the active layer at the permafrost monitoring sites	115
Figure 5-7: Comparison between frost table depths measured in 2015 or 2016 and 2018 at site (a) UB-1, (b) TIB-1, (c) HWY3-UB-1A and (d) ZF20-4a	117
Figure 5-8: Average apparent resistivity for permafrost monitoring sites between 2015 and 2018	118
Figure 5-9: Time-lapse electrical resistivity profiles presented as the percent (%) change in resistivity between 2015/16 and 2018 for (a) an unburned site (UB-1) and (b) a burned site (TIB-1) in the taiga shield ecozone, and (c) an unburned site (HWY3-UB-1A) and (d) a burned site (ZF20-4a) in the taiga plains ecozone	119
Figure 5-10: Average resistivity at 4 m depth in individual years compared to the average bulk density	120
Figure 5-11: Associations between gravimetric moisture content of organic soils in the active layers at burned sites and: (a) residual organic layer thickness, and (b) the portion of the organic layer lost during the fire	122

Figure 5-12: Associations between the portion of the organic layer lost during the fire and (a) the surface offset, and (b) the first quartile (Q1) of frost table depths, as well as (c) the relation of average resistivity at 4 m depth and the residual organic layer depth	123
Figure 5-13: Conditions at site ZF20-4a	124
Figure 5-14: Ground subsidence at ZF46-1b2	126
Figure 5-15: Changes in permafrost conditions at site ZF20-6b-LK between 2015 and 2019	127
Figure 5-16: Snow conditions at the monitoring sites between 2016-2019	129
Figure 5-17: Associations between: (a) annual snow-depth days and the nival offset; and (b) the site averaged snow-depth days (2016-2019) and the first quartile (Q1) of frost table depths ...	130
Figure 5-18: Average post-fire frost table depths compared to (a) years since fire and (b) the residual organic layer depth measured in this study (black) and in recent literature (white) at sites in the boreal forest	135
Figure 5-19: Changes in the first quartile (Q1) of frost table depths compared to average air temperature from 2015-2019	137

CHAPTER 1: INTRODUCTION

1.1 Introduction

Climate change is considered to be the most important and widespread environmental phenomenon of our immediate future, with the effects of global climate change projected to be most severe at high latitudes (IPCC, 2013). The IPCC (2013) shows that the observed climate warming of 0.8°C since the beginning of 20th century is strongly correlated to anthropogenic forcing compared to internal variability and other external drivers. Anthropogenic greenhouse gases in the atmosphere absorb long-wave radiation and reradiate it back to Earth, resulting in an overall warming effect. The Arctic is warming more than rest of the world, a phenomenon called Arctic Amplification (Serreze and Francis, 2006). Over the last century, average surface temperatures in the Arctic increased by $0.09^{\circ}\text{C}/\text{decade}$, while the rest of the Northern Hemisphere increased by $0.06^{\circ}\text{C}/\text{decade}$ (ACIA, 2005). Similarly, precipitation has increased by $1.4\%/\text{decade}$ from 1900-2003, with less falling as snow due to warmer temperatures (ACIA, 2005). This is expected to increase in the future, with 30-50% increases in winter precipitation over the Arctic Ocean, with more falling as rain instead of snow (AMAP, 2017). Further, there have been changes in the onset of snowmelt and delayed freeze-back in northern regions. In fact, recent climatic change is having an effect on all aspects of the cryosphere. There has been an overall reduction in sea ice extent, as well as less multi-year ice (Kwok, 2018; Serreze and Stroeve, 2015). Glaciers, ice caps, and ice sheets are melting and retreating (Benn and Evans, 2010; Sharp et al, 2014). The duration and extent of snow cover have generally decreased, particularly at high latitudes and elevations (AMAP, 2017), although snow depth has been increasing over northern Scandinavia and Russia (Bulygina et al., 2009). Similarly, the later freeze-up and earlier break-up of ice have combined to cause an overall reduction in the ice-cover season on arctic lakes and rivers (AMAP, 2017; Beltaos and Prowse, 2009; Benson et al., 2012; Prowse, 2011).

At many locations, permafrost temperatures have increased over last 30 years in response to observed climate warming (AMAP, 2017). In Canada, greater change has been observed in cold tundra sites ($1.2^{\circ}\text{C}/\text{decade}$) compared to warm ice-rich permafrost ($0.2^{\circ}\text{C}/\text{decade}$) (Burn and Kokelj, 2009; Smith et al., 2019). Permafrost along the Alaska and Mackenzie Highways has degraded since 1964, and where present it is thin and in organic soils (James et al., 2013; Kwong

and Gan, 1994). Predictions for the Arctic show that, by 2090, temperatures will have risen by 3°C to 6°C, and that precipitation, either as rainfall or snow cover, will increase (AMAP, 2017). These projected climatic changes are expected to continue to impact permafrost. For Canada, models that incorporate permafrost sensitivity based on variables such as climate, vegetation, and surficial geology project that by the end of the 21st century active layer thickness will increase, supra-permafrost taliks will develop, and permafrost may disappear entirely in southern regions (Smith and Burgess, 2004; Zhang et al., 2008).

One consequence of global climate change impacting permafrost landscapes is an increase in the frequency, severity, and extent of fires across the North American boreal forest (de Groot et al., 2013; Hanes et al., 2019; Jafarov et al., 2013; Kasischke et al., 2010; Wotton et al., 2017; Zhang et al., 2015). In Alaska, 50% more area was burned during the 2000s than in any decade in the previous 50 years (Kasischke et al., 2010). In Canada, the number of recorded fires has increased substantially in recent decades (Beck et al., 2011; Wotton et al., 2010). As global temperatures rise, there are more extended periods of drought, and larger and more severe fire years. The average fire return interval for western Canada is between 167-180 years, but is expected to decrease in the future (de Groot et al., 2013). Predictions are that the total area burned in Canada will increase by the end of the 21st century, with burned areas in the western boreal increasing by 3.5-5.5 times (Balshi et al., 2009; Flannigan et al., 2005) and the fire season will also lengthen (Stocks et al., 1998; Wang et al., 2015).

Fires have both immediate and long-term effects on the environment through changes in the surface energy balance (Kasischke and Stocks, 2000; Nossor et al., 2013). The radiation balance is the net radiative forcing at the ground surface (Oke, 1987), defined by:

$$Q^* = (K\downarrow - K\uparrow) + (L\downarrow - L\uparrow)$$

where Q^* is the net radiation, K is the short-wave radiation that is incident ($\downarrow$) or reflected ($\uparrow$), and L is the long-wave radiation that is incoming ($\downarrow$) or outgoing ($\uparrow$). This can be summarized as:

$$Q^* = K^* + L^*$$

where K^* is the net short-wave radiation and L^* is the net long-wave radiation. A simplified model of the energy exchange processes acting to dissipate the net radiation, otherwise known as the surface energy balance, is:

$$Q^* = Q_H + Q_E + Q_G$$

where Q_H , Q_E , and Q_G are the sensible, latent, and ground heat fluxes, respectively (Oke, 1987). Fires in the boreal forest alter the surface energy balance through a combination of factors. The dark, charred surface following a fire reduces the albedo, the surface absorbs more incident solar radiation, increasing K^* (Amiro, 2001; Rouse, 1976). There is also an increase in solar radiation incident on the surface due to the combustion of the canopy and reduced shading, further increasing K^* (Amiro, 2001). Fire enhances the loss of long-wave radiation from the surface (Amiro, 2001; Chambers and Chapin, 2002; Rouse, 1976). These combined factors lead to an overall reduction in Q^* (Amiro, 2001; Chambers and Chapin, 2002; Liu et al., 2005; Rouse, 1976).

The turbulent fluxes acting to dissipate the net radiation are also affected by fire, but, normally, more of the net radiation is partitioned into ground heat fluxes (Amiro, 2001; Chambers and Chapin, 2002). The ground heat flux increases due to a reduction in the above-ground biomass and organic mat thickness, which changes the thermal conductivity of the soil (Chambers and Chapin, 2002; Nossor et al., 2013). The sensible heat flux has been shown to both increase (Amiro et al., 1999; Rouse, 1976) and decrease (Amiro, 2001; Chambers and Chapin, 2002; Chambers et al., 2005; Liu et al., 2005) over burned areas, dependent on the pre-fire canopy characteristics, surface-atmosphere temperature gradients, and post-fire surface roughness. Overall, the latent heat flux appears to decrease following fire, as transpiring vegetation is destroyed (Amiro, 2001; Amiro et al., 1999; Chambers and Chapin, 2002; Chambers et al., 2005; Rouse, 1976). However, it is important to note that evaporation rates are variable, as they are controlled by post-fire and inter-annual fluctuations in soil moisture and climate, site specific characteristics, and burn severity (Amiro, 2001; Reed et al., 2016; Rouse, 1976). Soil moisture has been found to both increase (e.g. Fisher et al., 2016; Jiang et al., 2015) and decrease (e.g. Nossor et al., 2013; Swanson, 1996) following fire at permafrost sites, depending on characteristics like soil texture and topography. The combined factors affecting net radiation vary both diurnally and seasonally, where greatest post-fire effects occurring during the day and in the spring (Amiro, 2001; Liu et al., 2005; Rouse, 1976). Alterations in the energy balance persist for the first few years following fire, but return to pre-fire conditions after several decades as vegetation succession occurs (Amiro, 2001; Amiro et al., 2006; Liu et al., 2005; Rouse, 1976).

Forest fire is the most widespread form of natural surface disturbance to permafrost. Fires transform surface conditions, near-surface ground temperatures and soil moisture, the chemical properties of the soil, nutrient cycling, soil respiration rates, and species composition and other biological characteristics (Kasischke et al., 2010), and affect the water balance, and permafrost (Kasischke and Stocks, 2000). The combination of climate change and changes to the fire regime has the potential to increase permafrost degradation in the boreal forest (Yoshikawa et al., 2002; Shur and Jorgenson, 2007; Jorgenson et al., 2010; Jafarov et al., 2013).

Predicting how permafrost will change following fire in a warming climate is vital for residents and governments of northern regions, as well as for land and infrastructure managers. Forest fire is rarely taken into account in permafrost predictive modelling (e.g. Smith and Burgess, 2004; Zhang et al., 2008) even though fires have been shown to accelerate permafrost degradation (e.g. Gibson et al., 2018; Nossov et al., 2013; Smith et al., 2015). Modelling requires field validation across the range of ecosystems where permafrost is evolving. Furthermore, a large amount of permafrost in the discontinuous zone is thin, just below 0°C and is considered “ecosystem-protected”, meaning it is protected by intact forest vegetation and can persist where mean annual air temperature (MAAT) is as high as 2°C (Jorgenson et al., 2010). Following fire, it therefore becomes vulnerable to degradation and its loss has implications for hydrology, biogeochemistry, carbon cycling, and landscape stability.

1.2 Research objectives and thesis organization

The impetus for this thesis was the abnormal fire season of 2014 in the Northwest Territories (NWT), where 3.4 million ha of forested lands burned, making it the largest on record (1965 to present) in the NWT (Canadian Interagency Forest Fire Center, 2014). The fires affected the majority of NWT residents, including evacuations, health impacts caused by smoke inhalation (Kyle, 2015), and \$56.1 million spent fighting the fires (Hinchey, 2015). The 2014 fires presented a unique opportunity to access sites across a wide range of environmental conditions and levels of burn severity (Canadian Interagency Forest Fire Center, 2014; Walker et al., 2018). In particular, it enabled the creation of an experimental framework to examine how permafrost responds to natural surface disturbance in the sporadic and extensive discontinuous zones. During the summer of 2015, a multidisciplinary team of researchers came together to establish a network of permanent monitoring sites in fire scars from 2014, an outcome of which is Chapter 5 of this thesis.

Although the 2014 fires were unusually severe, and the impacts are expected to be widespread, long-term measurements are also required to address the full cycle of permafrost response. Historically, many variables which increased immediately following fire, such as active layer thickness and ground temperatures, returned to pre-fire levels after several decades (Gibson et al., 2018; Michaelides et al., 2019; Smith et al., 2015; Viereck et al., 2009). To extend the observations of the impacts of fire on permafrost in the region beyond what can be monitored directly during a Ph.D. program, a permafrost survey originally conducted in the 1960s (Brown, 1964) was reinvestigated to identify change which occurred over the intervening 55 years. This type of long-term information, particularly where pre-fire permafrost data is available, is extremely valuable to fully understand the effects of fire on permafrost as the ecosystem recovers in a warming climate.

This thesis aims to identify and address the knowledge gaps surrounding forest fire and permafrost interactions, by conducting in-situ measurement and analysis of permafrost conditions following fires occurring over the last half century, across a 650 km latitudinal transect spanning the discontinuous zones, from isolated patches (57.8°N) to extensive discontinuous permafrost (63.1°N), in northwestern Canada. The primary objective of this research is to understand how permafrost in the region, which until now has been sparsely investigated, is responding to fire in a warming climate.

We hypothesize that changes to permafrost depend on its pre-existing thermal regime, the severity of the burn, and ground ice content. Moderately burned areas may exhibit active layer deepening or the development of a supra-permafrost talik, with permafrost preserved at depth, whereas permafrost at temperatures close to 0°C and/or areas that are severely burned may completely degrade. This degradation will follow a latitudinal gradient, with northern sites showing little change, whereas permafrost at its most southerly margins may disappear entirely. Further, high ice content sites are expected to experience ground subsidence and the development of thermokarst. Excess ice contents are low across most of the study region (Figure 1-1; O'Neill et al., 2019; 2020), but high concentrations can be present locally, and where thermokarst occurs, it can dramatically affect the permafrost and surrounding landscape (Gibson et al., 2018).

The thermal and geomorphic changes initiated immediately following wildfire are expected to alter the hydrology and biogeochemistry of permafrost landscapes, including the release of previously frozen carbon (Bret-Harte et al., 2013; Jones et al., 2015; Li et al., 2019;

O'Donnell et al., 2011; Rocha and Shaver, 2011). Northern peatlands underlain by permafrost are particularly important because they store a huge amount of carbon (Schuur et al., 2008). In fact, intense burning of peatlands and permafrost thaw may turn these landscapes into carbon sources rather than sinks (Turetsky et al., 2002). However, the fate of carbon depends on its availability for decomposition. Peatlands are relatively young features with undegraded labile carbon, whereas frozen mineral soils can contain more old recalcitrant organics, but this also depends on the level of cryoturbation, paleoclimate, and disturbance history (Lacelle et al., 2019; Tarnocai et al., 2009).

This thesis contributes information about permafrost at burned sites to help establish how its changes could impact the ecosystem (or *vice versa*) and determine the potential for permafrost thaw to alter carbon stores. Furthermore, permafrost containing excess ice that thaws will undergo ground subsidence which can change forest succession pathways and be damaging to infrastructure (Burn, 1998). Modelling suggests that future climate warming, coupled with changes in fire regimes and thawing permafrost, will result in a reduction of carbon stocks (O'Donnell et al., 2011). This thesis provides information to clarify the timing and consequences of permafrost thaw.

This thesis is written in article format and each of the four main chapters represents a contribution to the goal of increasing the current state of knowledge on the impacts of fire on discontinuous permafrost in northwestern Canada. Three of these four chapters have been previously published. However, the additional reviews provided by the thesis committee members led to a number of minor typographical corrections and in a few instances, more substantive changes. Nevertheless, the major conclusions of these publications remained unchanged. The specific objectives addressed by each manuscript are indicated below.

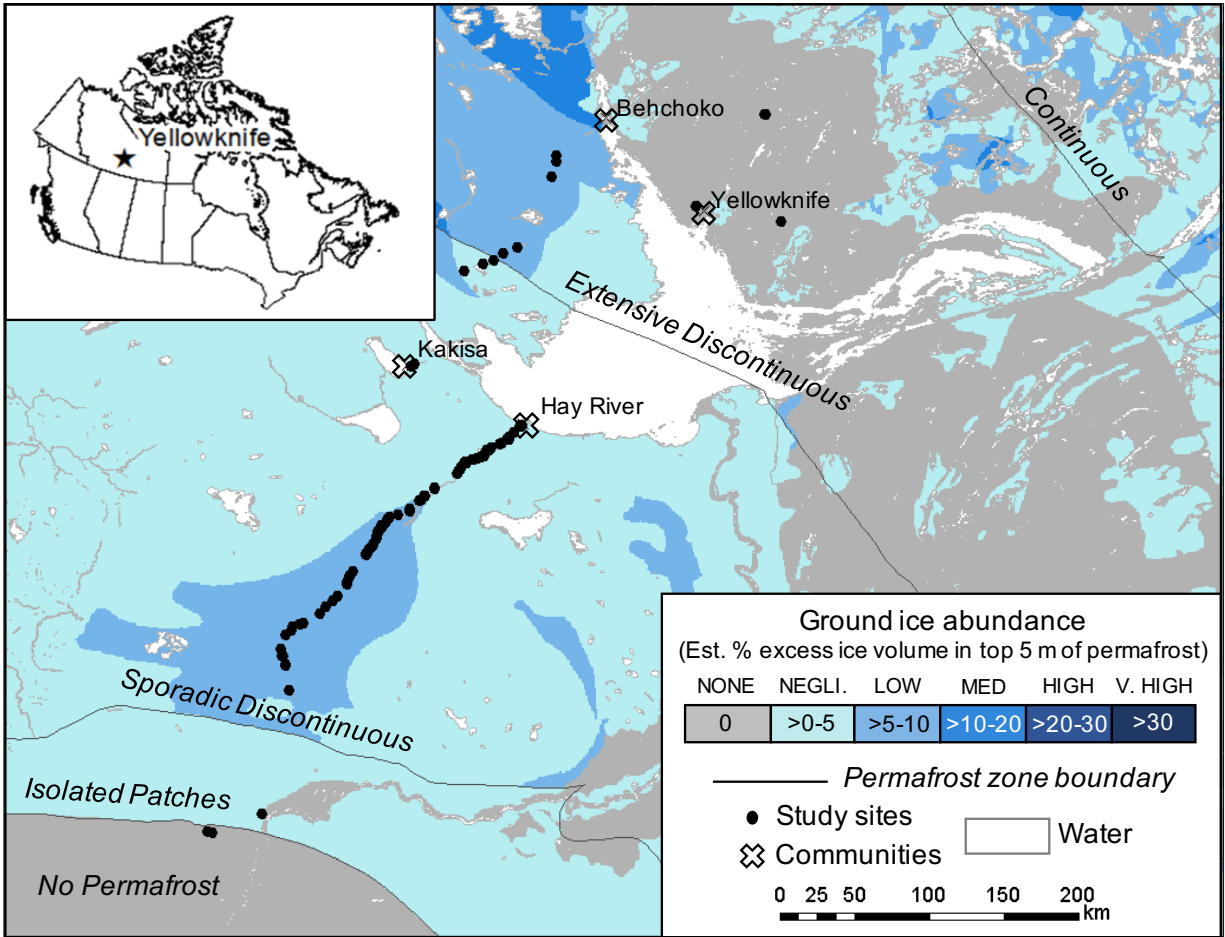


Figure 1-1: Map of the study region showing study sites in relation to modelled abundance of excess ice in the top 5 m of permafrost (O'Neill et al., 2019; 2020), permafrost zone boundaries, and major communities in the Northwest Territories. Note: negli. = negligible and v. high = very high. Permafrost zones are based on Heginbottom et al. (1995).

Objective 1: *Review recent literature and establish key knowledge gaps surrounding the impacts of fire on permafrost landscapes*

Chapter 2 is a literature review which provides details on recent (2010-2019) published research on the impacts of fire on permafrost landscapes across the northern hemisphere, including both the boreal forest and tundra. Published review papers are critical to the scientific process, as they provide platforms for developing new conceptual frameworks, reveal variations in the extant body of research, synthesize diverse results, and direct future research (Bem, 1995). In the past, permafrost has been mentioned as part of reviews focusing on ecological and biogeochemical impacts of fire (Douglas et al., 2014; Loranty et al., 2018). However, until this manuscript, there had been no comprehensive analysis of current knowledge surrounding post-

fire permafrost response, despite a large body of literature indicating the ongoing impacts, published both historically (e.g. Burn, 1998; Yoshikawa et al., 2002) and more recently (e.g. Brown et al., 2015; Zhang et al., 2015). This review synthesized results from more than 90 papers and identified key knowledge gaps, including: regional gaps of study within Canada; the lack of direct measurement of certain variables which affect post-fire permafrost dynamics, including soil moisture, snow depth, burn severity, and ground ice content and the quantification of subsidence and thermokarst development; the limited number of studies measuring impacts on permafrost at depth; and, the lack of investigations over a broad range of environmental conditions, particularly ground ice contents. Some of these knowledge gaps are addressed in the subsequent chapters of this thesis. Chapter 2 was published as part of a special issue of the journal *Permafrost and Periglacial Processes*.

Holloway, J. E., Lewkowicz, A. G., Douglas, T. A., Li, X., Turetsky, M. R., Baltzer, J. L., Jin, H. (2020). The impact of wildfire on permafrost landscapes: a review of recent advances and future prospects. *Permafrost and Periglacial Processes*, 31(3), 371-382. <https://doi.org/10.1002/ppp.2048>

Objective 2: *Establish long-term changes to permafrost following fire in a warming climate*

Chapter 3 represents a long-term analysis of how permafrost is responding to climate warming and fire in the sporadic discontinuous zone and isolated patches of permafrost. Long-term field studies of permafrost change are needed to fully understand post-fire permafrost response and to validate predictive models, but few are possible because of a lack of direct observations prior to the late 1970s, particularly for permafrost near its southernmost limit (Smith et al., 2010). To help fill this knowledge gap, we re-evaluated a permafrost survey originally undertaken in 1962. Similar studies have been conducted (James et al., 2013; Kwong and Gan, 1994), and are well cited as evidence of ongoing permafrost response to climate change, including in policy-directed reports such as those by the *Intergovernmental Panel on Climate Change* (IPCC, 2013). However, these studies were not able to critically address the role of fire in permafrost degradation over the long-term. Chapter 3 addresses this question by providing data on permafrost change over the last half century, which is needed to validate models of frozen ground response to climate change and fire in the discontinuous zone. This chapter was published in the journal *Permafrost and Periglacial Processes*.

Holloway, J. E., and Lewkowicz, A. G. (2020). Persistence of permafrost following 55 years of climate warming and forest fire in the discontinuous zone, Canada. *Permafrost and Periglacial Processes*, 31(1), 85-96. <https://doi.org/10.1002/ppp.2017>

Objective 3: *Determining the feasibility of using electrical resistivity tomography as a post-fire monitoring tool*

To date, most information on the impacts of fire on permafrost comes from near-surface measurements, neglecting permafrost conditions below 2 m. Electrical resistivity tomography (ERT) has been widely used as a means to delimit permafrost bodies to greater depths, due to the unique electrical properties of water that allow for differentiation between frozen and unfrozen ground (e.g. McClymont et al., 2013; Way et al., 2018). Similarly, repeat or continuous ERT measurements have been used to monitor permafrost change over time (Hilbich et al., 2011). Despite its use in other areas of permafrost study, ERT has rarely been used following fire (Douglas et al., 2016; Lewkowicz et al., 2011), and never repeatedly to monitor post-fire permafrost change over time.

One limitation of ERT is that the interpretation of images is partly subjective, due to the necessity for the user to choose a resistivity boundary between frozen and unfrozen soil (Hauck, 2013). This is especially challenging for soil with temperatures between -1 and 0°C, as there is a wide range of resistivity values and extensive overlap with those of unfrozen soil (Hoekstra et al., 1974). Further, the majority of studies utilizing ERT focus on relatively cold, ice-rich frozen ground, leading to the possible misinterpretation of warm permafrost bodies. Chapter 4 aims to address these specific issues, by comparing measurements made in the field and in the laboratory for different soils at a range of temperatures, and developing a novel approach for more accurate analysis of ERT images for discontinuous permafrost. Chapter 5 uses the thresholds between frozen and unfrozen soils established in Chapter 4 for interpretation of ERT results, and ultimately shows the effectiveness of repeat ERT surveys for monitoring permafrost change following fire. ERT was also employed in Chapter 3 as a means of elucidating permafrost conditions at depth after fires occurring over the last 50 years. Overall, the objective of this thesis is to assess the utility of ERT surveys as a post-fire permafrost monitoring tool. As mentioned above, Chapter 3 was published in the journal *Permafrost and Periglacial Processes*. Chapter 4 was published in the proceedings of the 8th Canadian Permafrost Conference.

Holloway, J., and Lewkowicz, A. G. (2019). Field and laboratory investigation of electrical resistivity-temperature relationships, southern Northwest Territories. In J.-P. Bilodeau, D.F. Nadeau, D. Fortier, and D. Conciatori (eds.), *Cold Regions Engineering 2019: Proceedings of the 18th International Conference on Cold Regions Engineering and 8th Canadian Permafrost Conference* (pp. 64-72). American Society of Civil Engineers, Reston, Virginia. <https://doi.org/10.1061/9780784482599>

Objective 4: *Establish short-term permafrost response following fire over a wide range of environmental conditions across the discontinuous zone*

Chapter 5 utilizes monitoring data collected over five years following the abnormally intense 2014 fire year in the NWT, Canada (Canadian Interagency Forest Fire Center, 2014). To date, most work studying the impact of fire on permafrost has been conducted in Alaska, with only a few studies in the NWT (e.g. Fisher et al., 2016; Gibson et al., 2018; Smith et al., 2015), even though some conditions, such as the limited topography across most of the taiga plains ecozone, are very different. Further, most studies focus on ice-rich permafrost, despite the fact that much frozen ground is low ice-content (Olefeldt et al., 2016; O'Neill et al., 2019; 2020). The 2014 fires burned an extensive portion of the discontinuous permafrost zones in the southern NWT, allowing access to sites with diverse environmental conditions (Walker et al., 2018b). Measurements were made of variables affecting post-fire permafrost response such as soil moisture, snow depth, burn severity, and ground ice content, including subsidence and thermokarst development. This manuscript provides important in-situ measurements of permafrost response to fire, which can be incorporated into models for future change. Chapter 5 is expected to be submitted to a refereed journal.

Holloway J.E., Lewkowicz A.G., Turetsky, M.R., Baltzer, J.L., Wolfe S.A., and Cumming S.G. Changes to permafrost in the 5 years after forest fire, southern Northwest Territories, Canada. To be submitted to a refereed journal.

Chapter 6 summarizes the research and results that make up the contents of the thesis, and presents overall conclusions, limitations, and suggestions for future work.

1.3 Authorship and co-authorship contributions

The candidate for this thesis is the primary author of the four manuscripts presented in Chapters 2, 3, 4, and 5. The overall research objectives, methodology for field data collection and approach for data analysis was established by the candidate in consultation with the thesis supervisor Dr. Antoni Lewkowicz. Field data collection was undertaken by the candidate and various field assistants over the course of the project. Each of the contributions was generated from the candidate's original ideas and writing, with guidance from Dr. Antoni Lewkowicz. Each of the four manuscripts presented in this thesis were improved by commentary from either research colleagues, conference attendees, anonymous reviewers and/or journal editors.

The authors of Chapter 2 are Jean Holloway, Dr. Antoni Lewkowicz, Dr. Thomas Douglas, Dr. Xiaoying Li, Dr. Merritt Turetsky, Dr. Jennifer Baltzer, and Dr. Huijun Jin. JH wrote the manuscript, with contributions from AL. Co-authors contributed knowledge on permafrost conditions in Alaska and China, as well as discussion of the contents of the paper, and manuscript editing.

The authors of Chapter 3 and Chapter 4 are Jean Holloway and Dr. Antoni Lewkowicz. JH and AL developed the methodology together. JH conducted field work, laboratory work, and data analysis. JH wrote the papers, with editing from AL.

The authors of Chapter 5 are Jean Holloway, Dr. Antoni Lewkowicz, Dr. Merritt Turetsky, Dr. Jennifer Baltzer, Dr. Stephen Wolfe, and Dr. Steve Cumming. All co-authors, save for JH and SW, were principal investigators on the project which acquired funds for field equipment. SC conducted the stratified random sampling used for site selection. MT, JB, and SW provided logistical support and guidance throughout the project. SW provided data from remote sites. JH conducted field and laboratory work, data analysis, and writing of the paper. All co-authors are expected to contribute edits to the manuscript before submission to an appropriate journal, but with the exception of AL, have yet to review the content.

1.4 References

- ACIA. (2005). *Arctic Climate Impact Assessment*. Cambridge University Press. 1020 pp.
- AMAP. (2017). *Snow, Water, Ice and Permafrost in the Arctic (SWIPA)*. Oslo, Norway: Arctic Monitoring and Assessment Programme.
- Amiro, B. D. (1999). BOREAS flight measurements of forest-fire effects on carbon dioxide and energy fluxes. *Agricultural and forest meteorology*, 96(4), 199-208. [https://doi.org/10.1016/s0168-1923\(99\)00050-7](https://doi.org/10.1016/s0168-1923(99)00050-7)
- Amiro, B. D. (2001). Paired-tower measurements of carbon and energy fluxes following disturbance in the boreal forest. *Global Change Biology*, 7(3), 253-268. <https://doi.org/10.1046/j.1365-2486.2001.00398.x>
- Amiro, B. D., Orchansky, A. L., Barr, A. G., Black, T. A., Chambers, S. D., Chapin III, F. S., Goulden, M. L., Litvak, M., Liu, H. P., McCaughey, J. H., McMillan, A., Randerson, J. T. (2006). The effect of post-fire stand age on the boreal forest energy balance. *Agricultural and Forest Meteorology*, 140(1-4), 41-50. <https://doi-org.proxy.bib.uottawa.ca/10.1016/j.agrformet.2006.02.014>
- Balshi, M. S., McGuire, A. D., Duffy, P., Flannigan, M., Walsh, J., Melillo, J. (2009). Assessing the response of area burned to changing climate in western boreal North America using a Multivariate Adaptive Regression Splines (MARS) approach. *Global Change Biology*, 15(3), 578-600. <https://doi.org/10.1111/j.1365-2486.2008.01679.x>
- Beck, P. S. A., Goetz, S. J., Mack, M. C., Alexander, H. D., Jin, Y., Randerson, J. T., Loranty, M. M. (2011). The impacts and implications of an intensifying fire regime on Alaskan boreal forest composition and albedo. *Global Change Biology*, 17(9), 2853–2866. <https://doi.org/10.1111/j.1365-2486.2011.02412.x>
- Beltaos, S. and Prowse, T. (2009). River-ice hydrology in a shrinking cryosphere. *Hydrological Processes*, 23(1), 122-144. <https://doi.org/10.1002/hyp.7165>
- Bem, D. J. (1995). Writing a review article for psychological bulletin. *Psychological Bulletin*, 118(2), 172–177.
- Benn, D., and Evans, D. J. (2010). Chapter 2: Snow, Ice and Climate. In *Glaciers and Glaciation*. Second Edition. Hodder Education. London, UK.
- Benson, B. J., Magnuson, J. J., Jensen, O. P., Card, V. M., Hodgkins, G., Korhonen, J., Livingstone, D. M., Stewart, K. M., Weyhenmeyer, G. A. and Granin, N. G. (2012). Extreme events, trends, and variability in Northern Hemisphere lake-ice phenology (1855–2005). *Climatic Change*, 112, 299-323. <https://doi.org/10.1007/s10584-011-0212-8>
- Bret-Harte, M. S., Mack, M. C., Shaver, G. R., Huebner, D. C., Johnston, M., Mojica, C. A., Pizano, C., and Reiskind, J. A. (2013). The response of Arctic vegetation and soils following an unusually severe tundra fire. *Philosophical Transactions of the Royal*

- Society B: Biological Sciences*, 368(1624), 20120490.
<https://doi.org/10.1098/rstb.2012.0490>
- Brown, R. J. E. (1964). *Permafrost investigations on the Mackenzie Highway in Alberta and Mackenzie district*. Ottawa, Canada, National Research Council Technical Paper No. 175.
- Brown, D. R. N., Jorgenson, M. T., Douglas, T. A., Vladimir, R. E., Kielland, K., Hiemstra, C., Euskirchen, E. S., and Ruess, R. W. (2015). Interactive effects of wildfire and climate on permafrost degradation in Alaskan lowland forests. *Journal of Geophysical Research - Biogeosciences*, 120(8), 1619-1637. <https://doi.org/10.1002/2015JG003033>
- Bulygina, O. N., Razuvaev, V. N., and Korshunova, N. N. (2009). Changes in snow cover over Northern Eurasia in the last few decades. *Environmental Research Letters*, 4(4), 045026. <https://doi.org/10.1088/1748-9326/4/4/045026>
- Burn, C. R. (1998). The response (1958-1997) of permafrost and near-surface ground temperatures to forest fire, Takhini River Valley, southern Yukon Territory. *Canadian Journal of Earth Science*, 35(2), 184-199. <https://doi.org/10.1139/e97-105>
- Burn, C. R. and Kokelj, S. V. (2009). The environment and permafrost of the Mackenzie Delta area. *Permafrost and Periglacial Processes*, 20(2), 83-105. <https://doi.org/10.1002/ppp.655>
- Canadian Interagency Forest Fire Center. (2014). *Situation Report – Sep 22, 2014*. <http://www.cifffc.ca/firewire/current.php?lang=en&date=20140922>
- Chambers, S. D., and Chapin III, F. S. (2002). Fire effects on surface-atmosphere energy exchange in Alaskan black spruce ecosystems: implications for feedbacks to regional climate. *Journal of Geophysical Research*, 107(D1), FFR 1-1 - FFR 1-17. <https://doi.org/10.1029/2001JD000530>
- Chambers S. D., Beringer, J., Randerson, J. T., Chapin III, F. S. (2005). Fire effects on net radiation and energy partitioning: contrasting responses of tundra and boreal forest ecosystems. *Journal of Geophysical Research*, 110(D9). <https://doi.org/10.1029/2004JD005299>
- de Groot, W. J., Flannigan, M. D., and Cantin, A. S. (2013). Climate change impacts on future boreal fire regimes. *Forest Ecology and Management*, 294, 35-44. <https://doi.org/10.1016/j.foreco.2012.09.027>
- Douglas, T. A., Jones, M. C., Hiemstra, C. A., and Arnold, J. R. (2014). Sources and sinks of carbon in boreal ecosystems of interior Alaska: A review. *Elementa Science of the Anthropocene*, 2, 000032. <http://doi.org/10.12952/journal.elementa.000032>
- Douglas, T. A., Jorgenson, M. T., Brown, D. R., Campbell, S. W., Hiemstra, C. A., Saari, S. P., Bjella, K., and Liljedahl, A. K. (2016). Degrading permafrost mapped with electrical

- resistivity tomography, airborne imagery and LiDAR, and seasonal thaw measurements. *Geophysics*, 81(1), 1JF-Z7. <https://doi.org/10.1190/geo2015-0149.1>
- Fisher, J. P., Estop-Aragónés, C., Thierry, A., Charman, D. J., Wolfe, S. A., Hartley, I. P., Murton, J. B., Williams, M., and Phoenix, G. K. (2016). The influence of vegetation and soil characteristics on active-layer thickness of permafrost soils in boreal forest. *Global Change Biology*, 22, 3127-3140. <https://doi.org/10.1111/gcb.13248>
- Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., and Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nature Communications*, 9, 3041. <https://doi.org/10.1038/s41467-018-05457-1>
- Hanes, C. C., Wang, X., Jain, P., Parisien, M., Little, J. M., and Flannigan, M. D. (2019). Fire-regime changes in Canada over the last half century. *Canadian Journal of Forest Research*, 49(3), 256-269. <https://doi.org/10.1139/cjfr-2018-0293>
- Hauck, C. (2013). New concepts in geophysical surveying and data interpretation for permafrost terrain. *Permafrost and Periglacial Processes*, 24(2), 131-137. <https://doi.org/10.1002/ppp.1774>
- Heginbottom, J. A., Dubreuil, M. A., and Harker, P. A. (1995). Canada – Permafrost. *National Atlas of Canada* (5th ed.). Natural Resources Canada.
- Hilbich, C., Fuss, C., and Hauck, C. (2011). Automated time-lapse ERT for improved process analysis and monitoring of frozen ground. *Permafrost and Periglacial Processes*, 22(4), 306-319. <https://doi.org/10.1002/ppp.732>
- Hinchey, G. (2015, May 6). 2014 N.W.T. fire season report: What you need to know. *CBC North*. <https://www.cbc.ca/news/canada/north/2014-n-w-t-fire-season-report-what-you-need-to-know-1.3061930#:~:text=In%202014%2C%20385%20fires%20burned,million%20hectares%20of%20forest%20land.&text=Despite%20the%20scale%20of%20last,no%20serious%20injuries%20or%20deaths>
- Hoekstra, P., Sellmann, P. V., and Delaney, A. J. (1974). *Airborne resistivity mapping of permafrost near Fairbanks, Alaska*. Cold Regions Research and Engineering Laboratory, Report 324.
- IPCC. (2013). Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. T. F. Stocker, D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex, and P. M. Midgley (Eds.). Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp. doi:10.1017/CBO9781107415324.
- Jafarov, E. E., Romanovsky, V. E., Genet, H., McGuire, A. D., and Marchenko, S. S. (2013). The effects of fire on the thermal stability of permafrost in lowland and upland black

- spruce forests of interior Alaska in a changing climate. *Environmental Research Letters*, 8(3), 035030. <http://dx.doi.org/10.1088/1748-9326/8/3/035030>
- James, M., Lewkowicz, A. G. Smith, S. L., and Miceli C. M. (2013). Multi-decadal degradation and persistence of permafrost in the Alaska Highway corridor, northwest Canada. *Environmental Research Letters* 8(4): 045013. <http://dx.doi.org/10.1088/1748-9326/8/4/045013>.
- Jiang, Y., Rocha, A. V., O'Donnell, J. A., Drysdale, J., Rastetter, E. B., Shaver, G. R., and Zhuang, Q. (2015). Contrasting soil thermal responses to fire in Alaskan tundra and boreal forest. *Journal of Geophysical Research: Earth Surface*, 120(2), 363-378. <http://dx.doi.org/10.1002/2014JF003180>
- Jones, B. M., Grosse, G., Arp, C. D., Miller, E., Liu, L., Hayes, D. J., and Larsen, C. F. (2015). Recent arctic tundra fire initiates widespread thermokarst development. *Scientific Reports*, 5, 15865. <https://doi.org/10.1038/srep15865>
- Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., O'Donnell, J., Schuur, E. A. G., Kanevskiy, M., and Marchenko, S. (2010). Resilience and vulnerability of permafrost to climate change. *Canadian Journal of Forest Research*, 40(7), 1219-1236. <https://doi.org/10.1139/X10-060>
- Kasischke, E. S., and Stocks, B. J. (2000). *Fire, Climate Change, and Carbon Cycling in the Boreal Forest*. Springer-Verlag.
- Kasischke, E. S., Verbyla, D. L., Rupp, T. S., McGuire, A. D., Murphy, K. A., Jandt, R., Barnes, J. L., Hoy, E. E., Duffy, P. A., Calef, M., and Turetsky, M. R. (2010). Alaska's changing fire regime - implications for the vulnerability of its boreal forests. *Canadian Journal of Forest Research*, 40(7), 1313-1324. <https://doi.org/10.1139/X10-098>
- Kwok, R. (2018). Arctic sea ice thickness, volume, and multiyear ice coverage: losses and coupled variability (1958–2018). *Environmental Research Letters*, 13(10), 105005. <https://doi.org/10.1088/1748-9326/aae3ec>
- Kwong, Y. J., and Gan, T. Y. (1994). Northward migration of permafrost along the Mackenzie Highway and climatic warming. *Climatic Change*, 26(4), 399-419. <https://doi.org/10.1007/BF01094404>
- Kyle, K. (2015, July 9). Ecology North gets \$130K to study health costs of forest fire smoke. *CBC North*. <https://www.cbc.ca/news/canada/north/ecology-north-gets-130k-to-study-health-costs-of-forest-fire-smoke-1.3144159>
- Lacelle, D., Fontaine, M., Pellerin, A., Kokelj, S. V., and Clark, I. D. (2019). Legacy of Holocene landscape changes on soil biogeochemistry: a perspective from paleo-active layers in northwestern Canada. *Journal of Geophysical Research: Biogeosciences*, 124(9), 2662-2679. <https://doi.org/10.1029/2018JG004916>

- Lewkowicz, A. G., Etzelmüller, B., and Smith, S. L. (2011). Characteristics of discontinuous permafrost based on ground temperature measurements and electrical resistivity tomography, southern Yukon, Canada. *Permafrost and Periglacial Processes*, 22(4), 320-342. <https://doi.org/10.1002/ppp.703>
- Li, X., Jin, H., He, R., Huang, Y., Wang, H., Luo, D., Jin, X., Lü, L., Wang, L., Li, W., Wei, C., Chang, X., Yang, S., and Yu, S. (2019). Effects of forest fires on the permafrost environment in the northern Da Xing'anling (Hinggan) mountains, Northeast China. *Permafrost and Periglacial Processes*, 30(3), 163-177. <https://doi.org/10.1002/ppp.2001>
- Liu, H. P., Randerson, J. T., Lindfors, J., Chapin III, F. S. (2005). Changes in the surface energy budget following fire in boreal ecosystems of interior Alaska: an annual perspective. *Journal of Geophysical Research*, 110(D13). <https://doi-org/10.1029/2004JD005158>
- Loranty, M. M., Abbott, B. W., Blok, D., Douglas, T. A., Epstein, H. E., Forbes, B. C., Jones, B. M., Kholodov, A. L., Kropp, H., Malhotra, A., Mamet, S. D., Myers-Smith, I. H., Natali, S. M., O'Donnell, J. A., Phoenix, G. K., Rocha, A. V., Sonnentag, O., Tape, K. D., and Walker, A. D. (2018). Reviews and syntheses: Changing ecosystem influences on soil thermal regimes in northern high-latitude permafrost regions. *Biogeosciences*, 15(17), 5287-5313. <https://doi.org/10.5194/bg-15-5287-2018>
- McClymont, A. F., Hayashi, M., Bentley, L. R., and Christensen, B. S. (2013). Geophysical imaging and thermal modeling of subsurface morphology and thaw evolution of discontinuous permafrost. *Journal of Geophysical Research: Earth Surface*, 118(3), 1826–1837. <https://doi.org/10.1002/jgrf.20114>
- Michaelides, R. J., Schaefer, K., Zebker, H. A., Parsekian, A., Liu, L., Chen, J., Natali, S., Ludwig, S., and Schaefer, S. R. (2019). Inference of the impact of wildfire on permafrost and active layer thickness in a discontinuous permafrost region using the remotely sensed active layer thickness (ReSALT) algorithm. *Environmental Research Letters*, 14(3), 035007. <https://doi.org/10.1088/1748-9326/aaf932>
- Nossov, D. R., Jorgenson, M. T., Kielland, K., and Kanevskiy, M. Z. (2013). Edaphic and microclimatic controls over permafrost response to fire in interior Alaska. *Environmental Research Letters*, 8(3), 035013. <https://doi.org/10.1088/1748-9326/8/3/035013>
- O'Donnell, J. A., Harden, J. W., McGuire, A. D., and Romanovsky, V. E. (2011). Exploring the sensitivity of soil carbon dynamics to climate change, fire disturbance and permafrost thaw in a black spruce ecosystem. *Biogeosciences*, 8, 1367-1382. <https://doi.org/10.5194/bg-8-1367-2011>
- Oke, T. R. (1987). *Boundary Layer Climates*. 2nd edition. Methuen, London.
- Olefeldt, D., Goswami, S., Grosse, G., Hayes, D., Hugelius, G., Kuhry, P., McGuire, A. D., Romanovsky, V. E., Sannel, A. B. K., Schuur, E. A. G., Turetsky, M. R. (2016).

- Circumpolar distribution and carbon storage of thermokarst landscapes. *Nature Communications*, 7, 13043. <https://doi.org/10.1038/ncomms13043>
- O'Neill, H. B., Wolfe, S. A., and Duchesne, C. (2019). New ground ice maps for Canada using a paleogeographic modelling approach. *Cryosphere*, 13, 753–773. <https://doi.org/10.5194/tc-13-753-2019>
- O'Neill, H. B., Wolfe, S. A., and Duchesne, C. (2020). *Ground ice map of Canada v.1*. Geological Survey of Canada, Open File 8713. In press.
- Prowse, T.D. (2011). Lake and river ice in Canada. In: French, H. and O. Slaymaker (Eds.), *Changing Cold Environments*. pp. 163-181. John Wiley and Sons.
- Reed, D. E., Ewers, B. E., Pendall, E., Frank, J., Kelly, R. (2016). Bark beetle-induced tree mortality alters stand energy budgets due to water budget changes. *Theoretical and Applied Climatology*, 131(1-2), 153-165. <https://doi.org/10.1007/s00704-016-1965-9>
- Rocha, A. V., and Shaver, G. R. (2011). Postfire energy exchange in arctic tundra: the importance and climatic implications of burn severity. *Global Change Biology*, 17(9), 2831–2841. <https://doi.org/10.1111/j.1365-2486.2011.02441.x>
- Rouse, W. R. (1976). Microclimatic changes accompanying burning in subarctic lichen woodland. *Arctic and Alpine Research*, 8(4), 357-376.
- Schuur, E. A. G., Bockheim, J., Canadell, J. G., Euskirchen, E., Field, C. B., Goryachkin, S. V., Hagemann, S., Kuhry, P., Lafleur, P. M., Lee, H., Mazhitova, G., Nelson, F. E., Rinke, A., Romanovsky, V. E., Shiklomanov, N., Tarnocai, C., Venevsky, S., Vogel, J. G., and Zimov, S. A. (2008). Vulnerability of permafrost carbon to climate change: implications for the global carbon cycle. *BioScience*, 58(8), 701–714. <https://doi.org/10.1641/B580807>
- Serreze, M. C., and Francis, J. A. (2006). The Arctic amplification debate. *Climatic Change*, 76(3-4), 241-264. <https://doi.org/10.1007/s10584-005-9017-y>
- Serreze, M. C., and Stroeve, J. (2015). Arctic sea ice trends, variability and implications for seasonal ice forecasting. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 373(2045), 20140159. <https://dx.doi.org/10.1098%2Frsta.2014.0159>
- Sharp, M., Burgess, D. O., Cawkwell, F., Copland, L., Davis, J. A., Dowdeswell, E. K., Dowdeswell, J. A., Gardner, A. S., Mair, D., Wang, L., and Williamson, S. N. (2014). Remote sensing of recent glacier changes in the Canadian Arctic. In G. J. Leonard, A. Kääb, J. S. Kargel, B. H. Raup, M. P. Bishop (eds.), *Global Land Ice Measurements from Space* (pp. 205-228). Springer Berlin Heidelberg.
- Shur, Y. L., and Jorgenson, M. T. (2007). Patterns of permafrost formation and degradation in relation to climate and ecosystems. *Permafrost and Periglacial Processes*, 18(1), 7-19. <https://doi.org/10.1002/ppp.582>

- Smith, S. L., and Burgess, M. (2004). *Sensitivity of permafrost to climate warming in Canada*. Bulletin of the Geological Survey of Canada (No. 579), Natural Resources Canada.
- Smith, S. L., Duchesne, C., and Lewkowicz, A. G. (2019). Tracking changes in permafrost thermal state in northern Canada. In J.-P. Bilodeau, D.F. Nadeau, D. Fortier, and D. Conciatori (eds.), *Cold Regions Engineering 2019: Proceedings of the 18th International Conference on Cold Regions Engineering and 8th Canadian Permafrost Conference* (pp. 64-72). American Society of Civil Engineers, Reston, Virginia. <https://doi.org/10.1061/9780784482599.077>
- Smith, S. L., Riseborough, D. W., and Bonnaventure, P. P. (2015). Eighteen-year record of forest fire effects on ground thermal regimes and permafrost in the central Mackenzie Valley, NWT, Canada. *Permafrost and Periglacial Processes*, 26(4), 289-303. <https://doi.org/10.1002/ppp.1849>
- Smith, S.L., Romanovsky, V.E., Lewkowicz, A.G., Burn, C.R., Allard, M., Clow, G.D., Yoshikawa, K., and Throop, J. (2010). Thermal state of permafrost in North America—a contribution to the International Polar Year. *Permafrost and Periglacial Processes*, 21(2), 117–135. <http://dx.doi.org/10.1002/ppp.690>
- Stocks, B.J., Fosberg, M.A., Lynham, T.J., Mearns, L., Wotton, B.M., Yang, Q., Jin, J.Z., Lawrence, K., Hartley, G.R., Mason, J.A. and McKenney, D.W. (1998). Climate change and forest fire potential in Russian and Canadian boreal forests. *Climatic Change*, 38, 1-13. <https://doi.org/10.1023/A:1005306001055>
- Swanson, D. K. (1996). Susceptibility of permafrost soils to deep thaw after forest fires in interior Alaska, U.S.A., and some ecologic implications. *Arctic and Alpine Research*, 28(2), 217-227.
- Tarnocai, C., Canadell, J. G., Schuur, E. A. G., Kuhry, P., Mazhitova, G., and Zimov, S. (2009). Soil organic carbon pools in the northern circumpolar permafrost region. *Global Biogeochemical Cycles*, 23(2). <https://doi.org/10.1029/2008GB003327>
- Turetsky, M. R., Wieder, R. K., Halsey, L., and Vitt, D. H. (2002). Current disturbance and the diminishing peatland carbon sink. *Geophysical Research Letters*, 29(11), 1-4. <https://doi.org/10.1029/2001GL014000>
- Viereck, L. A. (1982). Effects of fire and firelines on active layer thickness and soil temperatures in interior Alaska. In *Proceedings of the 4th Canadian Permafrost Conference, The Roger J. E. Brown Memorial Volume* (pp. 123-134). National Research Council of Canada.
- Viereck, L. A., Werdin-Pfisterer, N. R., Adams, P. C., and Yoshikawa, K. (2009). Effect of wildfire and fireline construction on the annual depth of thaw in a black spruce permafrost forest in interior Alaska: a 36-year record of recovery. In D. L. Kane and K.

- M. Hinkel (Eds.), *Proceedings of the 9th International Conference on Permafrost*, Institute of Northern Engineering, University of Alaska Fairbanks, Volume 2, 1845-1850.
- Walker, X. J., Rogers, B. M., Baltzer, J. L., Cumming, S. R., Day, N. J., Goetz, S., Johnstone, J. F., Schuur, E. A. G., Turetsky, M. R., and Mack, M. C. (2018b). Cross-scale controls on carbon emissions from boreal mega-fires of the Northwest Territories, Canada. *Global Change Biology*, 24(9), 4251-4265. <https://doi.org/10.1111/gcb.14287>
- Wang, X., Thompson, D. K., Marshall, G. A., Tymstra, C., Carr, R., and Flannigan, M. D. (2015). Increasing frequency of extreme fire weather in Canada with climate change. *Climatic Change*, 130, 573-586. <https://doi.org/10.1007/s10584-015-1375-5>
- Way, R. G., Lewkowicz, A. G., and Zhang, Y. (2018). Characteristics and fate of isolated permafrost patches in coastal Labrador, Canada. *The Cryosphere*, 12, 2667-2688. <https://doi.org/10.5194/tc-12-2667-2018>
- Wotton, B. M., Flannigan, M. D., and Marshall, G. A. (2017). Potential climate change impacts on fire intensity and key wildfire suppression thresholds in Canada. *Environmental Research Letters*, 12(9), 095003. <https://doi.org/10.1088/1748-9326/aa7e6e>
- Wotton, B. M., Nock, C. A., and Flannigan, M. D. (2010). Forest fire occurrence and climate change in Canada. *International Journal of Wildland Fire*, 19(3), 253-271. <https://doi.org/10.1071/WF09002>
- Yoshikawa, K., Bolton, W. R., Romanovsky, V. E., Fukuda, M., and Hinzman, L. D. (2002). Impacts of wildfire on the permafrost in the boreal forests of Interior Alaska. *Journal of Geophysical Research: Atmospheres*, 107(D1), FFR 4-1-FFR 4-14. <https://doi.org/10.1029/2001JD000438>
- Zhang, Y., Chen, W. J., and Riseborough, D. W. (2008). Transient projections of permafrost distribution in Canada during the 21st century under scenarios of climate change. *Global and Planetary Change*, 60(3-4), 443-456. <https://doi.org/10.1016/j.gloplacha.2007.05.003>
- Zhang, Y., Wolfe, S. A., Morse, P. D., Olthof, I., and Fraser, R. H. (2015). Spatiotemporal impacts of wildfire and climate warming on permafrost across a subarctic region, Canada. *Journal of Geophysical Research: Earth Surface*, 120(11), 2338-2356. <https://doi.org/10.1002/2015JF003679>

CHAPTER 2: IMPACT OF WILDFIRE ON PERMAFROST LANDSCAPES: A REVIEW OF RECENT ADVANCES AND FUTURE PROSPECTS

Abstract: Changes in the frequency and extent of wildfires are expected to lead to substantial and irreversible alterations to permafrost landscapes under a warming climate. Here we review recent publications (2010–2019) that advance our understanding of the effects of wildfire on surface and ground temperatures, on active layer thickness and, where permafrost is ice-rich, on ground subsidence and the development of thermokarst features. These thermal and geomorphic changes are initiated immediately following wildfire and alter the hydrology and biogeochemistry of permafrost landscapes, including the release of previously frozen carbon. In many locations, permafrost has been resilient, with key characteristics such as active layer thickness returning to pre-fire conditions after several decades. However, permafrost near its southern limit is losing this resiliency as a result of ongoing climate warming and increasingly common vegetation state changes. Shifts in fire return intervals, severity and extent are expected to alter the trajectories of wildfire impacts on permafrost, and to enlarge spatial impacts to more regularly include the burning of tundra areas. Modelling indicates some lowland boreal forest and tundra environments will remain resilient while uplands and areas with thin organic layers and dry soils will experience rapid and irreversible permafrost degradation. More work is needed to relate modelling to empirical studies, particularly incorporating dynamic variables such as soil moisture, snow and thermokarst development, and to identify post-fire permafrost responses for different landscape types and regions. Future progress requires further collaboration among geocryologists, ecologists, hydrologists, biogeochemists, modelers and remote sensing specialists.

2.1 Introduction

Wildfire is a normal, relatively infrequent (about 200 years for North America and 50 years in central Russia; Wotton et al., 2013), event in permafrost environments, particularly in boreal and subarctic forests but increasingly in tundra regions (Hanes et al., 2019; Kasischke et al., 2010; Rocha et al., 2012; Wang et al., 2015). Although perennially frozen ground can be resilient to this disturbance over long time scales, the growing frequency, severity, and extent of fires in the northern hemisphere have significantly impacted permafrost landscapes (de Groot et al., 2013; Hanes et al., 2019; Jafarov et al., 2013; Kasischke et al., 2010; Wotton et al., 2017;

Zhang et al., 2015). More frequent periods of drought, reduced snowpacks, more intense heat waves, and higher numbers of lightning strikes (Veraverbeke et al., 2017; Turetsky et al., 2011) have already resulted in larger fires and more severe fire seasons in the boreal, while fire season duration is expected to increase in tandem with a longer summer growing season (Wang et al., 2015). Tundra fires are rarer, but their frequency has also increased over the past several decades and is expected to double by the end of this century (Chipman et al., 2015; Rocha et al., 2012). The predicted increased frequency and magnitude of wildfires in combination with climate warming is expected to accelerate permafrost degradation and alter the post-fire recovery of permafrost landscapes (Douglas et al., 2016; Gibson et al., 2018; Jafarov et al., 2013; Zhang et al., 2015).

Fire has been shown to have both immediate and long-term effects on permafrost environments through changes in the surface energy balance, ground thermal and hydrological regimes, and soil and aquatic biogeochemistry (Bret-Harte et al., 2013; Jones et al., 2015; Li et al., 2019; O'Donnell et al., 2011b; Rocha and Shaver, 2011). The extensive literature, dating back half a century (e.g. Yoshikawa et al., 2002), generally examines the impacts of fire on permafrost using site-specific observations over relatively short time-scales, with some notable exceptions (Burn et al., 1998; Mackay, 1995). Recent publications assess permafrost responses at landscape scales and over decades, incorporating interactions among climate, vegetation, soil, and hydrology (Gibson et al., 2018; Zhang et al., 2015). Attention has been focused on determining which variables control permafrost system behaviour following fire (Fisher et al., 2016; Jafarov et al., 2013; Minsley et al., 2016; Nossor et al., 2013), the impacts of climate change on post-fire recovery of permafrost (Smith et al., 2015), and using modelling to predict the future resilience of perennially frozen ground (Jafarov et al., 2013; Jiang et al., 2015; Zhang et al., 2015).

This paper reviews publications on permafrost and fire, dating from 2010-2019 (98 cited and a further 64 listed in Appendix A2), reporting from within the discontinuous and continuous permafrost zones in Alaska, Canada, Russia, and China (Figure 2-1). We summarize the knowledge of wildfire impacts on permafrost landscapes, including an evaluation of the cumulative effects of wildfire and climate change, and identify gaps that require further investigation. We address post-fire (1) active layer and near-surface permafrost thermal regimes,

(2) thermokarst development, (3) permafrost recovery and resilience, and (4) changes to the hydrology and biogeochemistry of permafrost landscapes, including carbon cycling.

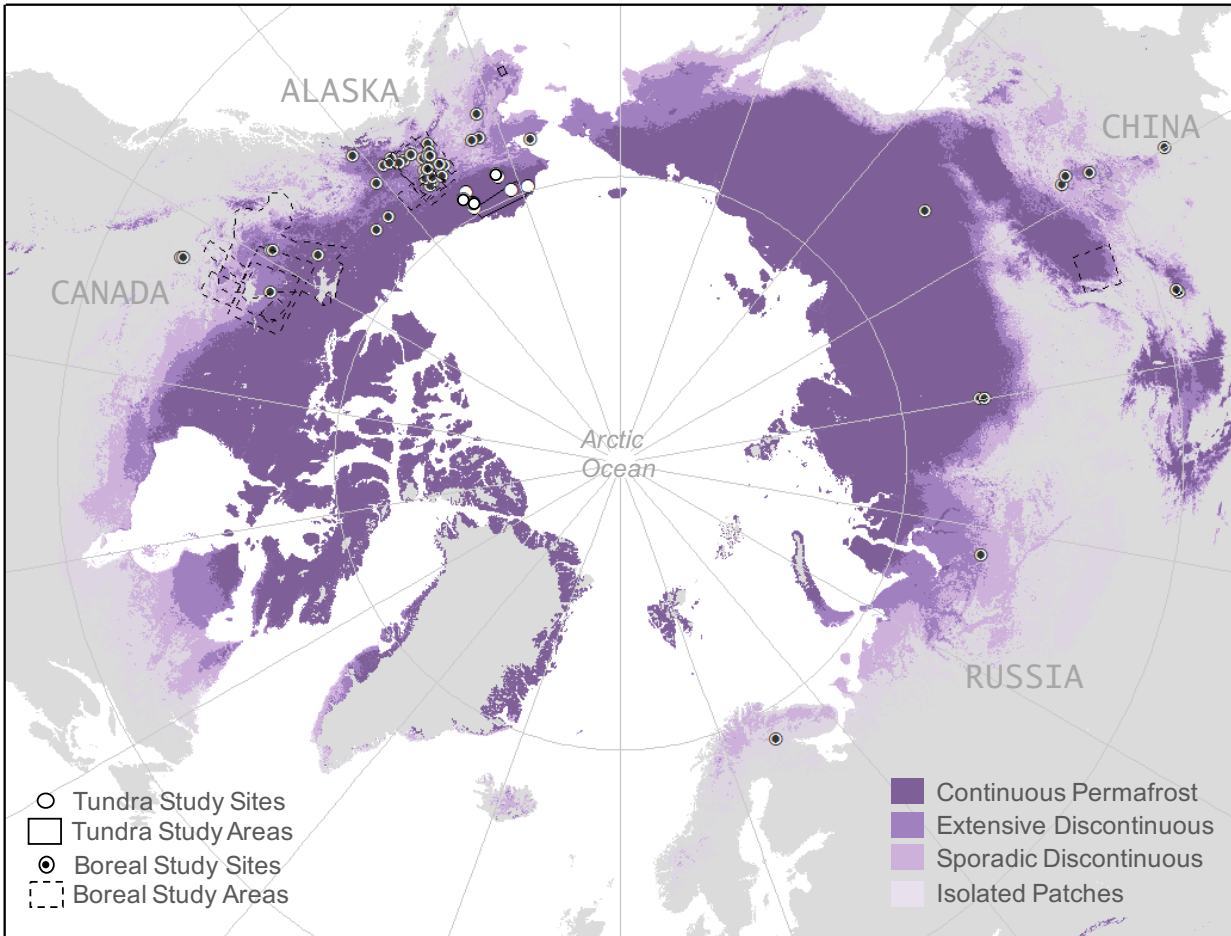


Figure 2-1: Locations of recent (2010-2019) studies on the impacts of wildfire on permafrost landscapes. Permafrost zones (Obu et al., 2018) applied to base map from Esri.

2.2 Post-fire changes to active layer thickness and the near-surface thermal regime

Wildfire and permafrost conditions both vary spatially and temporally. Pre-fire variations in current climate, recent climate change, topography, drainage, soil type, organic and moss layer thickness, and vegetation create spatial heterogeneity in ground temperatures and contribute to a patchwork of permafrost presence and absence in discontinuous permafrost zones. The direct impacts of fire in turn, can range from crown loss with little or no burning of the surface, to complete destruction of standing vegetation, as well as the surface organic and moss layers, resulting in exposure of the underlying mineral soil (Walker et al., 2018a). The amount of organic layer combustion and the post-fire residual organic layer thickness relate to the availability and composition of fuel, its bulk density and moisture content, and fire weather and behaviour (Benscoter et al., 2011; Johnstone et al., 2010; Walker et al., 2018a). Lowlands experience consistently lower burn severities than uplands, likely due to poor drainage and high soil moisture (Dillon et al., 2011; Turetsky et al., 2011). In these areas, however, smouldering is particularly important in organic layer consumption and this strongly affects the permafrost thermal regime (Benscoter et al., 2011). Permafrost itself can serve as a barrier to deep burning fires, as shown by late-season fires causing increased depth of burning in Alaskan forests except where permafrost maintains a shallow water table (Turetsky et al., 2011). Despite the superimposed spatial variability of permafrost and fire, however, consistent patterns of thermal response have been observed in different regions for similar landscapes, and we summarize these below.

Fire generally results in increased surface and soil temperatures for both boreal forest and tundra, beginning immediately following fire and lasting for several decades (Bai et al., 2013; Gibson et al., 2018; Jones et al., 2015; Jiang et al., 2015; Li et al., 2019; Nossov et al., 2013; O'Donnell et al., 2011b; Smith et al., 2015) (Table A1). Differences in mean annual ground surface temperatures of 1-7 °C have been observed at sites in Alaska, Canada, Mongolia, Siberia, and northeast China (Bai et al., 2013; Li et al., 2019; Nossov et al., 2013), with greater disparities in winter (Jiang et al., 2015; Smith et al., 2015). Within the soil column, differences between burned and unburned ground temperatures increase with burn severity, i.e., how much biomass is consumed (Bai et al., 2013; Li et al., 2019), but similar post-fire temperature increases may occur in both winter and summer (Jiang et al., 2015; Nossov et al., 2013; Smith et al., 2015). Surprisingly, lower ground temperatures were observed at some burned sites (Jiang et al., 2015;

Li et al., 2019), probably because of surface water pooling and subsequent increases in heat capacity, thermal conductivity, and evaporation.

Higher post-fire ground temperatures result in increased active layer thickness (ALT) in both boreal forest and tundra environments (Douglas et al., 2016; Fisher et al., 2016; Gibson et al., 2018; Jafarov et al., 2013; Jiang et al., 2015; Rocha et al., 2012; Smith et al., 2015; Zhang et al., 2015) (Figure 2-2; Table A1). The broad pattern is for a rapid increase in ALT in the years immediately following fire, reaching a maximum value after 5-10 years, followed by gradual recovery (Figure 2-2). However, progressive thaw can result in supra-permafrost talik development in both forested and tundra landscapes (Bret-Harte et al., 2013; Jafarov et al., 2013; Nossov et al., 2013; O'Neill et al., 2020). In northwestern Canadian peatlands, for example, talik extent over the last 30 years increased from 20% in unburned areas to 70-100% in burned locations (Gibson et al., 2018). As with ground temperatures, greater burn severity produces greater increases in post-fire thaw depths (Jafarov et al., 2013; Jiang et al., 2015; Li et al., 2019), but these can vary substantially, even over short distances, due to site-specific characteristics (Douglas et al., 2016). Remote sensing techniques, including airborne electromagnetic surveys and interferometric synthetic aperture radar, have recently been used to track post-fire changes in ALT over large areas of Alaskan tundra (Michaelides et al., 2019; Pastick et al., 2013). These techniques help elucidate regional heterogeneity in post-fire ALT.

Post-fire changes in the length of the seasonal thaw period have been observed in tundra and boreal landscapes in Alaska, Canada, Russia, and northeast China (Gibson et al., 2018; Jafarov et al., 2013; Li et al., 2019). Seasonal thaw started earlier in spring, and there was a delay in the start of active layer freeze-up (Jafarov et al., 2013). Thaw at a recently burned site in northwest Canada occurred more than two weeks earlier than at unburned sites, while pre-fire thaw processes resumed after about 30 years (Gibson et al., 2018).

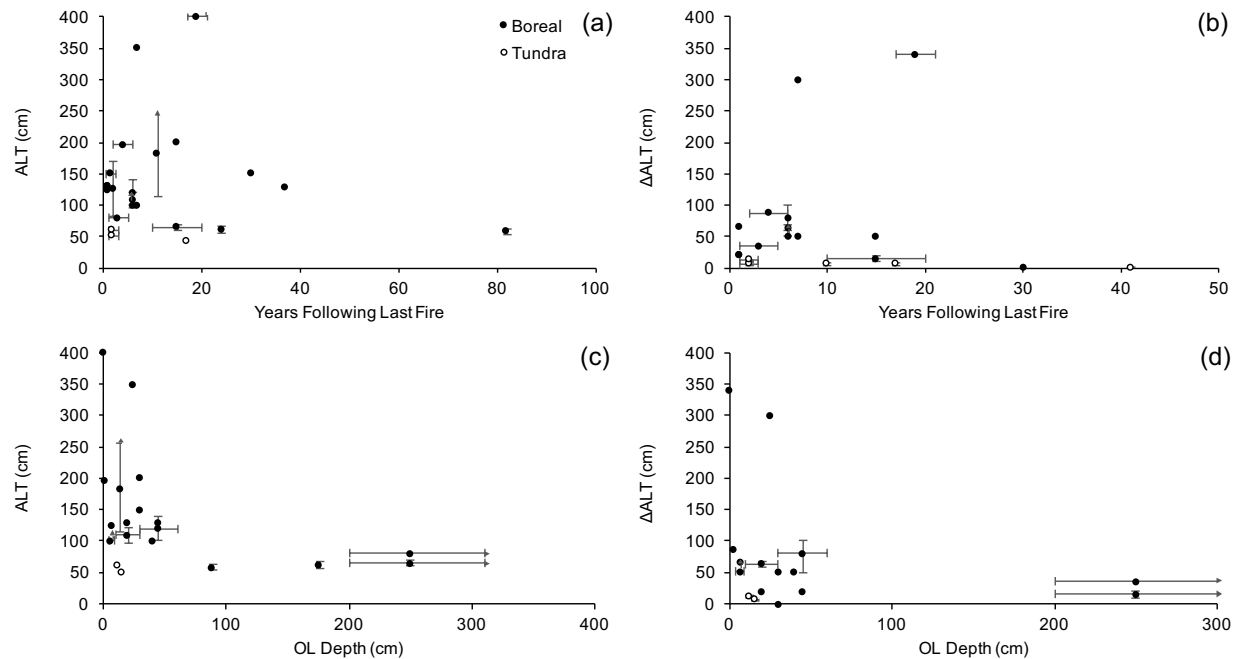


Figure 2-2: Observations of post-fire active layer thickness (ALT) from recent (2010-2019) literature for boreal and tundra sites (Douglas et al., 2016; Fisher et al., 2016; Gibson et al., 2018; Jafarov et al., 2013; Jiang et al., 2015; Li et al., 2019; Loranty et al., 2014; Nossor et al., 2013; Rocha and Shaver, 2011, 2012; Zhang et al., 2015;) (Table A1). (a) Post-fire ALT and (b) difference (Δ) in ALT between burned and unburned sites, both in relation to years since fire; (c) post-fire ALT and (d) difference (Δ) in ALT between burned and unburned sites, both in relation to residual organic layer (OL) depth including post-fire accumulation. Bars show the range of values (based on variable numbers of observations), dots are the means of the sample or the range, and arrows show where data were reported as being greater than or less than the value shown. Note: the ALT value of 400 cm and the Δ ALT of 350 cm likely include soil layers forming part of a supra-permafrost talik.

Winter snow cover affects the ground thermal regime in permafrost environments through the nival offset (Jorgenson et al., 2010). Following fire, the insulative properties of the snowpack in open, recovering forests are affected by reduced canopy interception that results in greater snow depths, greater snow redistribution, and compaction by wind (Burles and Boon, 2011; Jafarov et al., 2013). In addition, ablation rates are accelerated and snow disappears days to weeks earlier in a burned forest (Burles and Boon, 2011; Gleason et al., 2013) due to the decrease in snow albedo caused by carbon particles and charred woody debris, as well as increased solar radiation reaching the snow surface (Gleason et al., 2013). These changes in snow cover conditions and the timing of ablation must affect permafrost to some degree, but while lower post-fire interception losses are suspected of causing increased snow depths and ground temperatures at some sites in northwest Canada (Smith et al., 2015), there are few direct observations of these links.

At the landscape scale, total near-surface permafrost extent is expected to be affected by fire (Douglas et al., 2016; Zhang et al., 2015). Simulations for parts of northwestern Canada, for example, project greater reductions in near-surface permafrost extent for forested areas than for the tundra (Zhang et al., 2015). In contrast, enhanced rates of near-surface permafrost loss were not observed in the same region at forested sites where both 55 years of climate change and fire had occurred (Holloway and Lewkowicz, 2020). It is clear that predicting the response of permafrost to fire requires knowledge of both current permafrost extent and the degree of ecosystem protection. To map and forecast these changes, there is a pressing need to link permafrost degradation to biophysical characteristics that can be measured remotely.

2.3 Post-fire thermokarst development in ice-rich permafrost

The thermal changes described above that are brought on by fire lead to melting of ground ice in ice-rich permafrost. This may result in surface subsidence and the appearance of thermokarst features in both boreal and tundra environments (Bret-Harte et al., 2013; Brown et al., 2015; Douglas et al., 2016; Jones et al., 2015; Nossov et al., 2013; Minsley et al., 2016; Olefeldt et al., 2016) (Figure 2-3). These secondary impacts differ between ice-rich lowlands and sloping ice-rich terrain.

Paleo-reconstructions show that many permafrost peatlands have experienced cycles of post-fire thermokarst followed by autogenic succession of a peat layer and aggradation of permafrost (Jones et al., 2012). Post-fire thaw of ice-rich permafrost in these landscapes can be regarded as part of a cycle of disturbance and recovery, analogous to historical fire cycles (Zoltai, 1993). However, interactions between climate change and a changing fire regime may trigger permafrost thaw rates that exceed historical limits (Jones et al., 2013). Near the southern limit of permafrost in the Mackenzie Valley, northwest Canada, for example, fire is now responsible for an estimated 25% of thermokarst bog expansion (Gibson et al., 2018).



Figure 2-3: Photos of post-fire thermokarst development in areas where permafrost has high ice content: (a) aerial view of a site one-year post-fire in the Tanana Flats, Alaska. Note minor surface subsidence and inundation of surface water; (b) ground view of the area in (a) on the same date; (c) ground surface subsidence and water ponding one year post-fire at a site in the Taiga Shield, NWT, Canada; (d) water ponding due to thermokarst at a site in the northern Da Xing'anling Mountains, Northeast China, one year after fire; (e) active layer detachments in the Mackenzie Valley, Canada, one year after fire; and (f) active layer detachments in the Anaktuvuk River (Alaska) tundra burn, three years post-fire. Note: this figure was added after the original publication of this manuscript.

Near-surface massive ice features in permafrost, such as ice wedges, transition zone ice, relict buried ice, and zones of high ice content soils, are particularly vulnerable to melt following fire, leading to ground surface collapse in both the boreal forest (Brown et al., 2015) and tundra (Michaelides et al., 2019). Following the Anaktuvuk River tundra fire in Alaska, for example, 34% of the burned area experienced thaw and subsidence, compared to only 1% for unburned terrain (Jones et al., 2015). The variability in thermokarst development was linked to burn severity and ground ice content. Thermokarst development can lead to landscape wetting by promoting the transition from permafrost plateau forest to wetlands (Baltzer et al., 2014), or drying by causing ponds and lakes to drain (Jorgenson et al., 2013) or by lowering the water table (Walker et al., 2015). Thermokarst lake formation, as well as increases in lake size, are intensified by fire (Anderson et al., 2019; Manasypov et al., 2017).

Convective heat transfer by surface water may increase rates of lateral thaw (Lorantý et al., 2018; Neumann et al., 2019; O'Donnell et al., 2012). Additionally, ground subsidence promotes wind-driven increases in snow depths in landscape concavities, raising winter soil temperatures (Huscroft et al., 2004), particularly in tundra landscapes. Due to these processes and the loss of ecosystem protection, recent studies document an acceleration of ice-rich permafrost degradation following fire (Brown et al., 2015; Douglas et al., 2016; Minsley et al., 2016; Nossóv et al., 2013).

On slopes, post-fire active-layer detachments have been observed in the boreal forest (Huscroft et al., 2004; Lewkowicz et al., 2005; Lipovsky et al., 2005) (Figure 2-3e) and in the tundra (Jones et al., 2015) (Figure 2-3f). Rapid thaw of near-surface permafrost causes high porewater pressures to develop, which in turn reduce soil shear strength, leading to subsequent slope failure (Lewkowicz et al., 2005). Active-layer detachments were observed following fire in the discontinuous permafrost region of the Yukon, although geophysical results were ambiguous as to whether the loss of soil and vegetation within the scar zones caused supra-permafrost taliks to develop (Lewkowicz et al., 2011). In the Anaktuvuk River basin tundra fire, both active-layer detachments and retrogressive thaw slumps were triggered soon after the fire with both stabilizing approximately five years later (Jones et al., 2015). Although the spatial extent of these thermokarst features is usually quite small, they can disrupt infrastructure and have significant downstream impacts on aquatic ecosystems, and if they do not stabilize, they can lead to growing disturbance through time.

It is clear that fire can result in the development of ground subsidence and thermokarst in ice-rich permafrost settings. However, more work is needed to relate the response of permafrost to fire at the landscape scale, especially at interfaces between frozen and thawed material. Additionally, we need to improve our comprehension of how fire-induced thermokarst effects changes in hydrology, soils, and vegetation in order to fully assess the implications for ecosystem functioning.

2.4 Post-fire permafrost recovery and resilience

Permafrost change and recovery following wildfire depends on antecedent and post-fire conditions and the degree of disturbance (Jafarov et al., 2013; Nossov et al., 2013; Smith et al., 2015; Zhang et al., 2015) (Table 2-1). Permafrost has exhibited resilience to fire at many locations under extant climate conditions, with ALT, ground temperatures, and soil moisture returning to pre-fire levels after several decades (Gibson et al., 2018; Michaelides et al., 2019; Smith et al., 2015). For example, even though overlain by a 5-8 m thick supra-permafrost talik, permafrost at a Yukon site in the extensive discontinuous zone persisted to depths of >25 m more than a decade after fire (Northern Climate Exchange, 2013). Of the controlling variables, ground ice content is the most important: there are profound differences in permafrost response in settings where thermokarst develops versus those where there is little change in topography due to low or no excess ice. Permafrost can degrade rapidly where thermokarst develops and recovery in the past has occurred only over century to millennial time-scales (Zoltai, 1993). Whether recovery will occur in the future cannot be predicted in detail because current Earth System Models do not incorporate thermokarst dynamics. Regardless, historical patterns of permafrost recovery in both low and high ice content settings are expected to change as the climate continues to warm (Zhang et al., 2015) and a greater proportion of the remaining permafrost becomes dependent on ecosystem-protection (Shur et al., 2007). Compounding this, the structural elements that are associated with ecosystem-protected permafrost, e.g. conifer cover, moss carpets, or thick peat, are each projected to shift with climate warming (Dieleman et al., 2015; Mekonnen et al., 2019) but the degree to which this will impact permafrost is not known.

Table 2-1: Factors that control post-fire permafrost recovery (supported by the literature review).

Antecedent factors	Disturbance factors	Post-fire factors
• Ground ice content • Ground temperatures • Site topography • Site aspect • Site drainage and soil moisture • Site vegetation • Soil type • Organic layer thickness • Fire history • Magnitude of disequilibrium of ground temperatures with current climate	• Burn severity • Burn extent • Smouldering	• Residual organic layer thickness • Change in snowpack depth, density and duration • Drainage and soil moisture • Thermokarst development • Successional pathways and rates • Surface organic layer accumulation rate • Air temperature change during succession

Climate change affects the thermal response of burned and unburned permafrost sites unequally, with the former experiencing approximately 2 °C more warming over time at 1 m depth (Smith et al., 2015). While some burned sites are projected to recover after several decades, more extreme warming scenarios result in a longer recovery period (Zhang et al., 2015), and persistent effects on ALT and ground temperatures (Jafarov et al., 2013). For other sites, especially those that burned recently and where the ground is warmer, ALT is less likely to stabilize or recover if the climate has a warming trend (Brown et al., 2015; Zhang et al., 2016). Fire, therefore, renders permafrost more vulnerable to thaw under long-term climate change and *vice versa*, but modelling suggests that of the two, climate change will generally be the more important in the long-term (Zhang et al., 2015). However, an empirical study from north-western Canada showed stable permafrost under the spruce forest despite long-term climate warming, while degrading at nearby burned sites (Burn et al., 1998). These differences show the need to further validate modelling results. Fire could also delay permafrost loss in some peatlands and sparse forests if soil moisture increases post-fire but snow depth is unchanged due to low pre-fire stand density (Zhang et al., 2015).

Modelling shows that permafrost in poorly drained ecosystems, such as peatlands, tundra, and other lowland systems, tends to be more resilient over the long term and typically recovers more quickly to pre-fire levels (Genet et al., 2013; Jafarov et al., 2013; Zhang et al., 2015), due to thick soil organic layers both before and after wildfire, but only if these sites do not experience thermokarst development. Landscape position also contributes to post-fire permafrost recovery as topography affects microclimates, burn severity, drainage and soil moisture conditions, snow redistribution, and antecedent permafrost and vegetation conditions (Nossov et al., 2013; Smith et al., 2015; Walker et al., 2018a, 2019).

At the site level, several variables control how the frozen ground system behaves following fire. Residual organic layer thickness, changes in post-fire soil moisture, and vegetation recovery all affect active layer dynamics and permafrost recovery (Fisher et al., 2016; Jafarov et al., 2013; Minsley et al., 2016; Nossov et al., 2013). Identification of these variables in the literature is not new, but recent studies have focused on quantifying their relative importance. The role of these variables is reviewed below.

The thickness of the residual surface organic layer is generally the most important factor controlling post-fire active layer dynamics and the recovery of permafrost (Fisher et al., 2016; Jiang et al., 2015; Yi et al., 2010) (Figure 2-2). For example, removal of the entire organic layer is projected to result in thaw to a depth of 8 m within 30 years at an upland black spruce forest site in Alaska, even without considering climate warming (Jafarov et al., 2013). In contrast, a residual organic layer greater than 13 cm thick could prevent post-fire increase in ALT in the same region (Minsley et al., 2016; Yoshikawa et al., 2002).

Alterations to soil moisture and associated increases in thermal conductivity affect the post-fire thermal regime (Fisher et al., 2016; Jiang et al., 2015). Surface and deep soil moisture contents in the boreal forest are typically higher at burned sites than at unburned ones as a result of reduced evapotranspiration. These differences increase with burn severity (Bai et al., 2013; Fisher et al., 2016; Jiang et al., 2015) and can persist for extended periods, as demonstrated at an Alaskan site after 40 years (Jiang et al., 2015). These conditions coincide with the period of highest post-fire tree recruitment in northwest and north-central North American boreal forests (Johnstone et al., 2010). In contrast, burning of organic matter may create a hydrophobic layer on top of soils that lasts for several years (Kettridge et al., 2014) and this may impede infiltration and obstruct evaporation, confining higher moisture contents to the near-surface (Koch et al.,

2014). Post-fire reductions in soil moisture can occur at well-drained sites and in coarse-textured soils (Jorgenson et al., 2013; Li et al., 2019; Nossov et al., 2013; Swanson, 1996) and where post-fire increases in ALT lower water tables (Walker et al., 2015). On the other hand, little change in soil moisture was observed following a tundra fire in Alaska (Rocha and Shaver, 2011).

Higher post-fire moisture contents can promote permafrost persistence because wet soils encourage organic layer recovery and because ground ice is associated with higher latent heat contents (Jafarov et al., 2013; Nossov et al., 2013; Jiang et al., 2015). However, this applies only to permafrost sites with low to moderate ice contents as thaw settlement on ice-rich permafrost may result in surface ponding that triggers the development of thermokarst. Soil texture and ground ice content therefore shape the post-fire soil moisture regime, influencing the vulnerability of permafrost to degradation (Nossov et al., 2013). Burn severity, residual organic layer thickness, and post-fire permafrost recovery are all affected by soil moisture (Fisher et al., 2016; Walker et al., 2018a), so it is crucial to understand this variable, especially given predicted changes in precipitation regimes.

The rates and type of post-fire vegetation recovery (i.e. re-establishment of conifer or deciduous forest) also influence permafrost conditions, and depend on pre-fire conditions and burn severity (Table 2-1). Understory vegetation growth is rapid in the first few years following fire due to the warm ground surface and increased soil moisture, but progressively slows as there is more competition for moisture and sunlight (Van Cleve et al., 1983). Complete recovery of the canopy and understory vegetation to pre-burn conditions normally takes decades (Rocha et al., 2012). Permafrost in the North American boreal biome is commonly found beneath moist, black spruce or larch stands with thick organic layers, where post-fire moss recolonization is rapid, and mosses regain dominance after 2-3 decades, but only following low severity fires (Turetsky et al., 2010). The type of organic material also affects permafrost recovery, as feathermoss has a higher thermal conductivity than *Sphagnum* because the latter's conductivity does not change significantly after burning (Yoshikawa et al., 2002). Low-shrub and tussock-shrub are also common landscape types with different post-fire behaviours. Tussocks often survive burning, and shrubs resprout rapidly following fire (Jorgenson et al., 2013; Nossov et al., 2013; Zhang et al., 2015). Recovery of the organic layer at tundra sites following fire is similarly expected to

take several decades (Rocha and Shaver, 2011), although few long-term observations of post-fire permafrost recovery exist.

The rate of permafrost change is non-linear and slows five-years post-fire as vegetation recovers (Yoshikawa et al., 2002), but temperatures may still be several degrees warmer than at unburned sites some four decades later (Gibson et al., 2018; Jiang et al., 2015; Smith et al., 2015). Burn severity affects recovery times with severely burned sites taking decades longer than those moderately burned in both tundra and boreal environments (Jafarov et al., 2016; Jiang et al., 2015; Zhang et al., 2015). This is particularly true for ecosystems with thin organic layers and soils with low to moderate moisture content, where permafrost degradation can be rapid and irreversible. Similarly, severe burning of sites can result in vegetation state changes (Johnstone et al., 2010; Mekonnen et al., 2019) that may alter the local surface energy balance through changes in albedo and snowpack properties and accumulation of organic soils, but the degree to which permafrost is impacted by these transitions requires further investigation. Ecological research illustrates the impacts of re-burning on organic layer thickness (Hoy et al., 2016; Walker et al., 2019), but there has been little work to directly measure the impacts of re-burns on permafrost. The implications of the expected shortened fire return intervals require investigation at a full range of sites.

Based on the literature, we have compiled a series of simplified trajectories for the post-fire response of permafrost (Figures 2-4 and 2-5). The trajectory followed depends on all the site-specific factors listed in Table 2-1, but antecedent ground ice content is considered pre-eminent. At low ice-content sites, the potential trajectories summarized in Figure 2-4 have been outlined previously (Zhang et al., 2015). For sites with moderate to high ice-content the post-fire trajectory depends on landscape position (Figure 2-5).

In ice-rich lowlands (Figure 2-5a), the active layer deepens and the surface settles differentially due to thaw of ground ice in the uppermost permafrost layers. Ponding may occur due to poor drainage, enhancing rates of top-down and lateral permafrost thaw (Figure 2-3a to d). Over time, bog development or thaw lake development creates a very deep talik or completely thaws permafrost. Vegetation succession causes long-term changes in ground temperatures beneath bogs but permafrost is unlikely to re-develop there due to intervening projected climate change. Permafrost also does not re-develop beneath thaw lakes unless they drain and then only if conditions remain cold enough despite intervening climate warming. In contrast, permafrost

can persist in peat plateaus and the active layer can return to pre-fire conditions subject to climate change related increases.

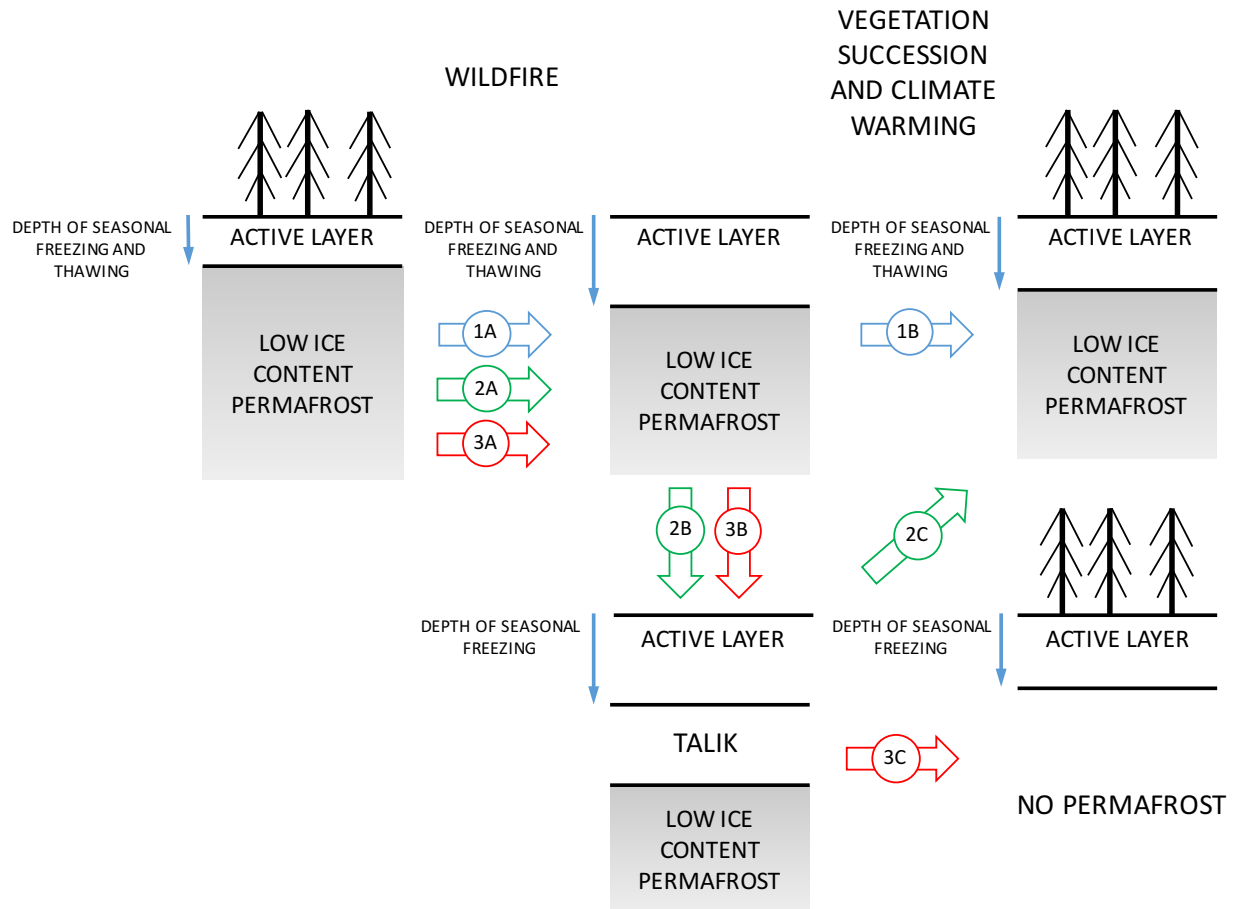


Figure 2-4: Conceptual diagram showing the potential post-fire trajectories for permafrost without excess ice and hence limited or no surface subsidence during thaw. In trajectory 1 (blue arrows), the active layer becomes thicker, permafrost temperatures increase but remain $<0^{\circ}\text{C}$ and succession causes ground temperatures and the active layer to eventually recover but not to pre-fire values due to locally warming climate. In trajectory 2 (green arrows), a talik forms as thaw of the upper permafrost exceeds maximum seasonal freezing, permafrost is maintained at depth with higher temperatures but still $<0^{\circ}\text{C}$, and succession causes ground temperatures to eventually recover but not to pre-fire values due to intervening climate change. In trajectory 3 (red arrows), a talik forms and progressively deepens as permafrost thaws from above and below (not shown) until it has entirely degraded; succession causes ground temperatures to recover partially but insufficiently to re-develop permafrost due to intervening climate change. The successional sequence may also be altered by the impact of permafrost loss on drainage. The trajectory followed depends on the site-specific factors listed in Table 2-1.

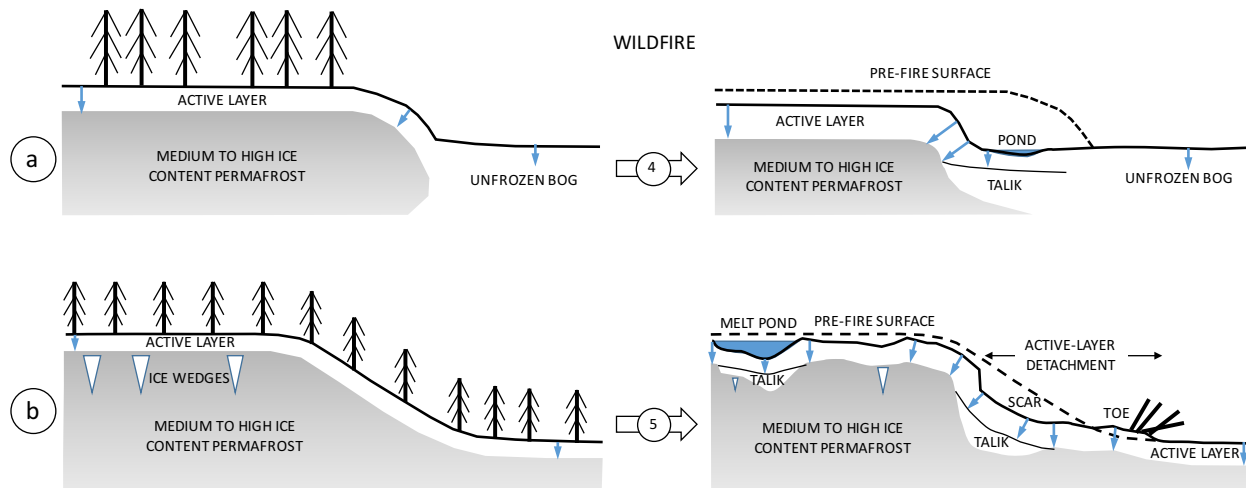


Figure 2-5: Conceptual diagram showing the potential post-fire trajectories for permafrost containing excess ice and subject to surface subsidence during thaw. (a) In trajectory 4, a peat plateau in a poorly-drained lowland experiences both surface settlement and lateral loss following wildfire. A talik develops at the edge of the unfrozen bog, often with a shallow pond, as permafrost thaws rapidly due to convective heat transfer. (b) In trajectory 5, rapid thaw following wildfire of a slope underlain by ice-rich permafrost leads to the formation of an active layer detachment (ALD) or retrogressive thaw slump (not shown). A talik may develop beneath the bare scar in the ALD while slight permafrost aggradation may occur beneath the toe. On the crest of the slope, vertical settlement dominates with the formation of depressions filled by melt water in particularly ice-rich locations. In both (a) and (b) subsequent steps in the trajectories can follow paths 1b, 2c or 3c in Figure 2-4. The trajectory followed depends on the site-specific factors listed in Table 2-1. Note: blue arrows show the depth of seasonal freezing and a talik is present where this does not reach the depth of the permafrost table.

In uplands with moderate to high ice-content, the active layer deepens beneath slopes after fire (Figure 2-5b), and gullies or landslides (active-layer detachments and retrogressive thaw slumps) can be triggered by rapid thaw of high ice-content soil layers and/or massive ice features (Figure 2-3e and f). The uppermost permafrost layers then thaw further beneath landslide scars due to the absence of overlying soil and vegetation, resulting in a greater ALT or talik formation. In these ice-rich settings, slope is extremely important in facilitating post-thaw water runoff, which largely prevents the impoundment that could cause further permafrost thaw. On hillcrests, the deepening active layer following fire results in differential thaw settlement and melt ponds may develop over bodies of ground ice. The subsequent trajectories for both hillslopes and hillcrests could end in complete thaw or partial recovery, depending on location, as with low ice-content sites (see Figure 2-4).

2.5 Hydrological and biogeochemical impacts of wildfire in permafrost landscapes

The impacts of wildfires on hydrological and biogeochemical processes in permafrost environments vary regionally and from site to site (Kopp et al., 2017; Tank et al., 2018, 2020),

but some broad generalizations are possible. However, most recent findings are based on investigations of permafrost in boreal forest environments so it remains to be determined if they apply to the tundra biome.

The hydrologic regimes of permafrost catchments are affected by loss of surface vegetation and subsequent changes in snow (see above) that can increase the volume and magnitude of runoff (Gleason et al., 2013), resulting in earlier spring freshet and increased peak flows (Owens et al., 2013; Seibert et al., 2010). Overland flow and stormflow runoff increase mainly because the water retention capacity of the surface organic layer is diminished, resulting in a decline in baseflow (Kopp et al., 2017). Over the longer term, increases in ALT alter water storage (Kopp et al., 2017; Nossov et al., 2013) while talik development creates perennial pathways between surface and groundwater systems (Connon et al., 2014). These changes affect the routing of water through the landscape and amplify the effects of fire, with groundwater propagating laterally, transferring advective heat from burned to unburned regions. Shallow closed taliks are also common in burned and thawing peatlands (Gibson et al., 2018), but instead of connecting water they isolate it, promoting further talik and bog expansion. In coarse-grained upland areas (which are often extensive), post-fire thaw may stop permafrost from acting as an aquitard and promote vertical drainage instead of downslope supra-permafrost flow.

Fire disturbance affects the biogeochemistry and soil microbial community of permafrost landscapes (Kawahigashi et al., 2011; Kong et al., 2014; Taş et al., 2014). Direct impacts include reducing nutrient uptake by vegetation, which changes runoff solute sources, reactions, and transport mechanisms, as well as chemically and physically altering the surface organic layer (Burd et al., 2018; Verma and Jayakumar, 2012).

Generally, fire increases soil pH (Day et al., 2019; Kawahigashi et al., 2011; Taş et al., 2014) and reduces dissolved organic carbon (DOC), C, N, and C/N ratios (Guo et al., 2011; Kawahigashi et al., 2011; Taş et al., 2014), NO_3^- , PO_4^{3-} , and P (Gu et al., 2010). At a tundra site in Alaska, short-term increases in the availability of inorganic N and P occurred, but plant-available N was not significantly different between burned and unburned sites four years after fire (Bret-Harte et al., 2013). Post-fire nitrogen mineralization is particularly important as this affects permafrost through vegetation recovery and the local energy balance.

Wildfire directly alters the soil microbial community composition and metabolic potential, and reduces the potential for respiration (Köster et al., 2015; Zhou et al., 2019). It also

indirectly increases rates of microbial activity by increasing soil temperature and moisture (O'Donnell et al., 2011b). These changes in turn affect C, N, and P cycling relating to the mineralization of organic material to inorganic forms (Zhou et al., 2019). Post-fire increases in pH of the soil reduce the richness and diversity of mycorrhizal fungi (Day et al., 2019; Sun et al., 2015; Zhou et al., 2019), and can shift soil microbial communities from fungal-dominated to bacterial-dominated (Aaltonen et al., 2019; Zhou et al., 2019). Post-fire, the soil microbial community structure and soil enzyme activity recover dynamically (Zhou et al., 2019) but permafrost thaw and changes to soil moisture may alter microbial activity and fungal abundance. Long-term studies are needed to examine this question, particularly in the context of changing fire regimes (Day et al., 2019).

Post-fire increases in ALT and temperature, and talik development, affect watershed biogeochemical processes by altering the solute chemistry of soil water, groundwater, stream water and lake water, as well as changing the routing of water and its biogeochemical components (Anderson et al., 2019; Manasypov et al., 2017; Tank et al., 2018). Post-fire permafrost thaw increases nutrient mineralization rates and releases previously frozen porewater containing solutes, nutrients, organics and contaminants, into the surrounding aquatic environment (Gordon et al., 2016; Schuster et al., 2011; Tank et al., 2018). Deep thaw and talik development enable the flow of these biogeochemical components year-round (Allison et al., 2011; Tank et al., 2018). Large post-fire increases in runoff N, P, and DOC have generally been observed (Koch et al., 2014; Tank et al., 2018). Post-fire permafrost thaw may continue to release previously immobilized solutes, resulting in increased basin exports (Brown et al., 2015; Gordon et al., 2016; Schuster et al., 2011), but the influence of fire on stream chemistry declines as the system returns to pre-fire conditions. Aquatic ecosystems seem to be relatively resilient to wildfire in their catchments (Lewis et al., 2014).

Fire has long-term consequences on the ecosystem C balance, as increased soil temperatures and ALT make previously frozen organic matter available for decomposition by soil microbes that release C in gaseous form into the atmosphere (Genet et al., 2013; Hu et al., 2012; O'Donnell et al., 2011a, 2011b; Turetsky et al., 2011; Taş et al., 2014; Walz et al., 2017; Zhou et al., 2019). In the past, decomposition was reduced as the ALT returned to pre-fire conditions because organic matter was returned to a frozen state (O'Donnell et al., 2011b; Yi et al., 2010). Under a warmer climate, thawed permafrost will take longer to recover (Zhang et al.,

2015), if at all, meaning labile organic carbon stocks will be vulnerable to decomposition for longer periods. Furthermore, high-severity fires result in a net loss of carbon from the system (O'Donnell et al., 2011a). At the same time, post-fire decreases in microbial biomass, changes to the microbial community structure, and losses of labile carbon, may limit decomposition post-fire (Aaltonen et al., 2019; Köster et al., 2015; Sun et al., 2015; Taş et al., 2014; Zhou et al., 2019). Post-fire carbon accumulation rates have been identified as having the potential to offset losses, but net primary productivity would have to increase substantially (O'Donnell et al., 2011a; Yi et al., 2010). Therefore, it appears likely that increases in the frequency and magnitude of fire, and subsequent changes to permafrost, will result in increased carbon loss (O'Donnell et al., 2011a). Regional carbon pool estimates need to be improved in order to verify this, particularly for the Russian sub-arctic (Hugelius et al., 2014).

Some analyses show that the terrestrial carbon sinks may be shifting to sources (McGuire et al., 2010; O'Donnell et al., 2011a; Walker et al., 2019), but modelling is needed to fully evaluate the combined effects of fire, hydrology, permafrost and changing climate, including snow dynamics (O'Donnell et al., 2011a; Walker et al., 2019). Thermokarst development can cause rapid thaw of large volumes of carbon to considerable depths (Nossov et al., 2013) while thickening of the active layer affects only the top few cm of permafrost and does so incrementally. Despite evidence that wildfire can initiate thermokarst and accelerates its expansion (Brown et al., 2015; Douglas et al., 2016; Gibson et al., 2018), current thermal models predicting the impacts of fire on permafrost do not incorporate these processes, and it is not known whether carbon cycling behaviour differs between burned sites with thermokarst and unburned degrading sites.

2.6 Discussion: Priorities for future research

Many of the papers published between 2010-2019 examined the impacts of wildfire on permafrost environments, focusing on edaphic or ecological aspects. As a result, there remains much to be learned in relation to permafrost itself.

A majority of the recent publications examined wildfire impacts on boreal forest sites in Alaska, resulting in regional gaps for parts of Canada, Siberia, and Mongolia (Figure 2-1). We now have a deep understanding of how Alaskan permafrost sites react and it is reasonable to assume that post-fire permafrost trajectories in the western Canadian boreal forest are similar, for

areas that have comparable MAATs (Table A1). However, verification is needed that these conclusions apply to the larch-dominated forests in Siberia, Mongolia, and northeast China. The eastern Canadian boreal forest in Quebec and Labrador also remains little studied, and, since this region has very different physiographic, climatological, and fire regimes compared to western Canada, it is likely that post-fire permafrost behaviour may be dissimilar.

The impacts of fire on discontinuous permafrost in the boreal biome are better understood than fire effects on continuous permafrost in the tundra, despite fires becoming more common in these environments, as with the Anaktuvuk River fire and western Greenland fires of 2017. Almost all tundra fire research has been carried out in Alaska (Figure 2-1) even though the burned area of the Russian tundra is an order of magnitude larger. Knowledge in all regions is incomplete as no recent publications include long-term monitoring of current rates of recovery of ALT and surface and ground temperatures. Both the frequency and magnitude of fires and their impacts may be changing because of the combination of climate warming and the expansion of shrubs into tundra areas (Elmendorf et al., 2012; Myers-Smith et al., 2018).

Regional upscaling from site-specific results within the boreal and tundra biomes remains challenging due to the heterogeneity of landscapes (and even of individual burn scars), especially in relation to ground temperatures and ground ice content. Overcoming this challenge requires the use of remote sensing, which in turn depends on improvements in methods to measure and validate biophysical characteristics linked to post-fire permafrost degradation.

Most empirical studies of fire impacts on lowland permafrost demonstrate rapid change at high ice-content sites with thermokarst (Baltzer et al., 2014; Gibson et al., 2018), while modelling studies typically make predictions of slower change for low ice-content sites without thermokarst. In the past, permafrost recovery after thermokarst formation and the development of bogs and thaw lakes occurred on century to millennial time scales (Zoltai, 1993), but we are not using the proper tools to assess current recovery rates. Paleoreconstructions cannot adequately answer such questions and Earth System Models do not fully integrate thermokarst development. Additional empirical studies are needed to assess the impacts of loss of low ice-content permafrost on the hydrology, carbon fluxes, and biogeochemistry of uplands with coarse-grained soils, since extensive swaths of the permafrost region are of this landscape type (Olefeldt et al., 2016).

Several key variables in the permafrost system that are affected by fire have been little studied to date. These include snow depth, soil moisture and organic layer recovery, all of which affect permafrost presence and characteristics. In addition, within the fire system, we know less about ground fires in the north and smouldering combustion, even though the latter has been identified as a dominant form of combustion in peat soils and can result in greater depths of burn. Similarly, more empirical data are needed to characterize ground ice conditions and quantify where and how much ground subsidence and thermokarst development takes place following fire, and how this affects the hydrologic regime, forest succession, and complex feedbacks amongst these components.

There are few investigations of the impact of wildfire on hydrology in permafrost landscapes, particularly in medium to large basins (Owens et al., 2013), while incorporation of dynamic post-fire landscape behaviour, for instance regarding soil moisture or thermokarst dynamics, into hydrological and environmental models remains a challenge (Hinzman et al., 2003). In general, uplands in boreal and Arctic regions are expected to become drier as the climate warms, and lowlands, wetter, but the combined impact of these trends and changing fire regimes on permafrost resilience has not been fully assessed. Models suggest resiliency of lowland boreal permafrost and tundra landscapes following fire (Jafarov et al., 2013; Zhang et al., 2015), but empirical evidence is not as clear on this point (Brown et al., 2015; Gibson et al., 2018; Jones et al., 2015). The discrepancy between the two types of result is at least partly because ground ice needs better representation in models. Calibrations of the latter would be improved by additional field-based inputs, particularly if dynamic variables such as soil moisture, snow, and vegetation are incorporated.

Regional modelling of future permafrost extent needs to incorporate the impacts of fire. Empirical and modelling studies have documented that fire induces permafrost change and that fire and climate change are inducing vegetation state changes, such as expansion of deciduous forest, but these results have been produced by different disciplines and scientists that rarely share methods or conferences. Cross-disciplinary collaborative work among geocryologists, ecologists, biogeochemists, modellers and experts in remote sensing is needed to fully understand the impacts on permafrost of the many complex post-fire ground, vegetation and drainage changes.

2.7 Conclusions

The recently published literature examining the impacts of wildfire on permafrost has focused on understanding landscape response, monitoring long-term changes and utilizing modelling techniques to evaluate regional changes and predict the impacts of future climate warming. Fire causes immediate changes to the permafrost, increasing ALT, surface and ground temperatures, and soil moisture. Post-fire thermokarst development and ground subsidence may occur where permafrost is ice-rich. Increases in ALT, talik and thermokarst development release previously frozen porewater containing solutes, including carbon, and create new hydrological pathways, allowing the year-round transport of biochemical components.

Many permafrost landscapes showed past resilience, returning to near pre-fire conditions after several decades, unless thaw of near-surface ice-rich permafrost triggered thermokarst formation. However, modelling suggests that this will change as the climate warms and fire regimes shift, so that permafrost at burned sites will recover more slowly or not at all. It appears that colder lowland boreal forest and tundra environments in general will remain resilient to fire in the medium term. In contrast, permafrost near its southern margin depends on ecosystem protection and will degrade entirely after fire, following trajectories that may be altered by vegetation state changes.

The complexity of the relationships among wildfire, climate, vegetation, soils, hydrology and ground thermal state means that no single discipline can fully evaluate the impacts of fire on permafrost. We conclude that additional work is needed to relate models to empirical studies, to identify patterns from different permafrost landscape types and regions, and to bridge results from the ecological, geocryological, biogeochemical, remote sensing and modelling communities.

2.8 References

- Aaltonen, H., Palviainen, M., Zhou, X., Köster, E., Berninger, F., Pumpanen, J., and Köster, K. (2019). Temperature sensitivity of soil organic matter decomposition after forest fire in Canadian permafrost region. *Journal of Environmental Management*, 241, 637-644. <https://doi.org/10.1016/j.jenvman.2019.02.130>
- Allison, S. D., and Treseder, K. K. (2011). Climate change feedbacks to microbial decomposition in boreal soils. *Fungal Ecology*, 4(6), 362-374. <https://doi.org/10.1016/j.funeco.2011.01.003>
- Anderson, L., Edwards, M., Shapley, M. D., Finney, B. P., and Langdon, C. (2019). Holocene thermokarst lake dynamics in northern Interior Alaska: the interplay of climate, fire, and subsurface hydrology. *Frontiers in Earth Science*, 7(53), 1-22. <https://doi.org/10.3389/feart.2019.00053>
- Baltzer, J. L., Veness, T., Chasmer, L. E., Sniderhan, A. E., and Quinton, W. L. (2014). Forests on thawing permafrost: fragmentation, edge effects, and net forest loss. *Global Change Biology*, 20(3), 824-834. <https://doi.org/10.1111/gcb.12349>
- Bai, Q., Zhou, M., Qi, Q., Meng, R., Cui, X., and Wen, D. (2013). The changes of the frozen soil temperature and humidity of the forest fire forest in the Great Hinggan Mountains. *Journal of Arid Land Resources and Environment*, 5(27), 69-75.
- Benscoter, B. W., Thompson, D. K., Waddington, J. M., Flannigan, M. D., Wotton, B. M., De Groot, W. J., and Turetsky, M. R. (2011). Interactive effects of vegetation, soil moisture and bulk density on depth of burning of thick organic soils. *International Journal of Wildland Fire*, 20(3), 418-429. <https://doi.org/10.1071/WF08183>
- Bret-Harte, M. S., Mack, M. C., Shaver, G. R., Huebner, D. C., Johnston, M., Mojica, C. A., Pizano, C., and Reiskind, J. A. (2013). The response of Arctic vegetation and soils following an unusually severe tundra fire. *Philosophical Transactions of the Royal Society B: Biological Sciences*, 368(1624), 20120490. <https://doi.org/10.1098/rstb.2012.0490>
- Brown, D. R. N., Jorgenson, M. T., Douglas, T. A., Vladimir, R. E., Kielland, K., Hiemstra, C., Euskirchen, E. S., and Ruess, R. W. (2015). Interactive effects of wildfire and climate on permafrost degradation in Alaskan lowland forests. *Journal of Geophysical Research - Biogeosciences*, 120(8), 1619-1637. <https://doi.org/10.1002/2015JG003033>
- Burd, K., Tank, S. E., Dion, N., Quinon, W. L., Spence, C., Tanentzap, A. J., and Olefeldt, D. (2018). Seasonal shifts in export of DOC and nutrients from burned and unburned peatland-rich catchments, Northwest Territories, Canada. *Hydrology and Earth System Science*, 22(8), 4455-4472. <https://doi.org/10.5194/hess-22-4455-2018>

- Burles, K., and Boon, S. (2011). Snowmelt energy balance in a burned forest plot, Crowsnest Pass, Alberta, Canada. *Hydrological Processes*, 25(19), 3012-3029. <https://doi.org/10.1002/hyp.8067>
- Burn, C. R. (1998). The response (1958-1997) of permafrost and near-surface ground temperatures to forest fire, Takhini River Valley, southern Yukon Territory. *Canadian Journal of Earth Science*, 35(2), 184-199. <https://doi.org/10.1139/e97-105>
- Chipman, M. L., Hudspeth, V., Higuera, P. E., Duffy, P. A., Kelly, R., Oswald, W. W., and Hu, F. S. (2015). Spatiotemporal patterns of tundra fires: Late-Quaternary charcoal records from Alaska. *Biogeosciences*, 12, 4017-4027. <https://doi.org/10.5194/bg-12-4017-2015>
- Connon, R. F., Quinton, W. L., Craig, J. R., and Hayashi, M. (2014). Changing hydrologic connectivity due to permafrost thaw in the lower Liard River valley, NWT, Canada. *Hydrological Processes*, 28(14), 4163-4178. <https://doi.org/10.1002/hyp.10206>
- Day, N. J., Dunfield, K. E., Johnstone, J. F., Mack, M. C., Turetsky, M. R., Walker, X. J., White, A. L., and Baltzer, J. L. (2019). Wildfire severity reduces richness and alters composition of soil fungal communities in boreal forests of western Canada. *Global Change Biology*, 25(7), 2310-2324. <https://doi.org/10.1111/gcb.14641>
- de Groot, W. J., Flannigan, M. D., and Cantin, A. S. (2013). Climate change impacts on future boreal fire regimes. *Forest Ecology and Management*, 294, 35-44. <https://doi.org/10.1016/j.foreco.2012.09.027>
- Dieleman, C. M., Branfireun, B. A., McLaughlin, J. W., and Lindo, Z. (2015). Climate change drives a shift in peatland ecosystem plant community: Implications for ecosystem function and stability. *Global Change Biology*, 21(1), 388-395. <https://doi.org/10.1111/gcb.12643>
- Dillon, G. K., Holden, Z. A., Morgan, P., Crimmins, M. A., Heyerdahl, E. K., and Luce, C. H. (2011). Both topography and climate affected forest and woodland burn severity in two regions of the western US, 1984 to 2006. *Ecosphere*, 2(12), 1-33. <https://doi.org/10.1890/ES11-00271.1>
- Douglas, T. A., Jones, M. C., Hiemstra, C. A., and Arnold, J. R. (2014). Sources and sinks of carbon in boreal ecosystems of interior Alaska: A review. *Elementa Science of the Anthropocene*, 2, 000032. <http://doi.org/10.12952/journal.elementa.000032>
- Douglas, T. A., Jorgenson, M. T., Brown, D. R., Campbell, S. W., Hiemstra, C. A., Saari, S. P., Bjella, K., and Liljedahl, A. K. (2016). Degrading permafrost mapped with electrical resistivity tomography, airborne imagery and LiDAR, and seasonal thaw measurements. *Geophysics*, 81(1), 1JF-Z7. <https://doi.org/10.1190/geo2015-0149.1>
- Elmendorf, S. C., Henry, G. H. R., Hollister, R. D., Björk, R. G., Boulanger-Lapointe, N., Cooper, E. J., Cornelissen, J. H. C., Day, T. A., Dorrepaal, E., Elumeeva, T. G., Gill, M., Gould, W. A., Harte, J., Hik, D. S., Hofgaard, A., Johnson, D. R., Johnston, J. F.,

- Jónsdóttir, I. S., Jorgenson, J. C., ... Wipf, S. (2012). Plot-scale evidence of tundra vegetation change and links to recent summer warming. *Nature Climate Change*, 2, 453-457. <https://doi.org/10.1038/nclimate1465>
- Fisher, J. P., Estop-Aragónés, C., Thierry, A., Charman, D. J., Wolfe, S. A., Hartley, I. P., Murton, J. B., Williams, M., and Phoenix, G. K. (2016). The influence of vegetation and soil characteristics on active-layer thickness of permafrost soils in boreal forest. *Global Change Biology*, 22, 3127-3140. <https://doi.org/10.1111/gcb.13248>
- Genet, H., McGuire, A. D., Barrett, K., Breen, A., Euskirchen, E. S., Johnstone, J. F., Kasischke, E. S., Melvin, A. M., Bennett, A., Mack, M. C., Rupp, T. S., Schuur, A. E. G., Turetsky, M. R., and Yuan, F. (2013). Modelling the effects of fire severity and climate warming on active layer thickness and soil carbon storage of black spruce forests across the landscape in interior Alaska. *Environmental Research Letters*, 8, 045016. <http://dx.doi.org/10.1088/1748-9326/8/4/045016>
- Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., and Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nature Communications*, 9, 3041. <https://doi.org/10.1038/s41467-018-05457-1>
- Gleason, K. E., Nolin, A. W., and Roth, T. R. (2013). Charred forests increase snowmelt: effects of burned woody debris and incoming solar radiation on snow ablation. *Geophysical Research Letters*, 40(17), 4654-4661. <https://doi.org/10.1002/grl.50896>
- Gordon, J., Quinton, W., Branfireun, B. A., and Olefeldt, D. (2016). Mercury and methylmercury biogeochemistry in a thawing permafrost wetland complex, Northwest Territories, Canada. *Hydrological Processes*, 30(20), 3627-3638. <https://doi.org/10.1002/hyp.10911>
- Gu, H., Jin, J., Cheng, X., Wang, E., Zhou, Y., and Chai, Y. (2010). The long-term impacts on chemical properties of *Larix gmelini* forest on the northern slope of Greater Hinggan Mountains from a forest fire of varying fire intensity. *Journal of Natural Resources*, 25(7), 1114-1121 (in Chinese).
- Guo, A., Guo, Y., and Cui, X. (2011). Effects of different intensities of fire disturbances on soil nutrients in a *Pinus massoniana* forest in the Greater Xing'an Mountain. *Journal of Northeast Forestry University*, 39(5), 69-71.
- Hanes, C. C., Wang, X., Jain, P., Parisien, M., Little, J. M., and Flannigan, M. D. (2019). Fire-regime changes in Canada over the last half century. *Canadian Journal of Forest Research*, 49(3), 256-269. <https://doi.org/10.1139/cjfr-2018-0293>
- Hinzman, L. D., Fukuda, M., Sandberg, D. V., Chapin, F. S., and Dash, D. (2003). FROSTFIRE: an experimental approach to predicting the climate feedbacks from the changing boreal fire regime. *Journal of Geophysical Research Atmospheres*, 108(D1), FFR 9-1-FFR 9-6. <https://doi.org/10.1029/2001JD000415>

- Holloway, J. E., and Lewkowicz, A. G. (2020). Persistence of permafrost following 55 years of climate warming and forest fire in the discontinuous zone, Canada. *Permafrost and Periglacial Processes*, 31(1), 85-96. <https://doi.org/10.1002/ppp.2017>
- Hoy, E. E., Turetsky, M. R., and Kasischke, E. S. (2016). More frequent burning increases vulnerability of Alaskan boreal black spruce forests. *Environmental Research Letters*, 11, 095001. <http://dx.doi.org/10.1088/1748-9326/11/9/095001>
- Hu, H., Wei, S., and Sun, L. (2012). Estimating carbon emissions from forest fires during 2001 to 2010 in Daxing'anling Mountain. *Shengtai Xuebao/Acta Ecologica Sinica*, 32(17), 5373-5386. <http://dx.doi.org/10.5846/stxb201109091324>
- Hugelius, G., Strauss, J., Zubrzycki, S., Harden, J.W., Schuur, E. A. G., Ping, C.-L., Schirmer, L., Grosse, G., Michaelson, G. J., Koven, C. D., O'Donnell, J. A., Elberling, B., Mishra, U., Camill, P., Yu, Z., and Palmtag, J. (2014). Estimated stocks of circumpolar permafrost carbon with quantified uncertainty ranges and identified data gaps. *Biogeosciences*, 11, 6573-6593. <https://doi.org/10.5194/bg-11-6573-2014>
- Huscroft, C. A., Lipovsky, P. S., and Bond, J. D. (2004). *A regional characterization of landslides in the Alaska Highway corridor, Yukon*. Whitehorse, Canada: Yukon Geological Survey, Open File 18.
- Jafarov, E. E., Romanovsky, V. E., Genet, H., McGuire, A. D., and Marchenko, S. S. (2013). The effects of fire on the thermal stability of permafrost in lowland and upland black spruce forests of interior Alaska in a changing climate. *Environmental Research Letters*, 8(3), 035030. <http://dx.doi.org/10.1088/1748-9326/8/3/035030>
- Jiang, Y., Rocha, A. V., O'Donnell, J. A., Drysdale, J., Rastetter, E. B., Shaver, G. R., and Zhuang, Q. (2015). Contrasting soil thermal responses to fire in Alaskan tundra and boreal forest. *Journal of Geophysical Research: Earth Surface*, 120(2), 363-378. <http://dx.doi.org/10.1002/2014JF003180>
- Johnstone, J. F., Hollingsworth, T. N., Chapin, F. S., and Mack, M. C. (2010). Changes in fire regime break the legacy lock on successional trajectories in Alaskan boreal forest. *Global Change Biology*, 16(4), 1281-1295. <https://doi.org/10.1111/j.1365-2486.2009.02051.x>
- Jones, B. M., Grosse, G., Arp, C. D., Miller, E., Liu, L., Hayes, D. J., and Larsen, C. F. (2015). Recent arctic tundra fire initiates widespread thermokarst development. *Scientific Reports*, 5, 15865. <https://doi.org/10.1038/srep15865>
- Jones, M. C., Booth, R. K., Yu, Z., and Ferry, P. (2013). A 2200-year record of permafrost dynamics and carbon cycling in a collapse-scar bog, interior Alaska. *Ecosystems*, 16, 1-19. <https://doi.org/10.1007/s10021-012-9592-5>
- Jones, M. C., Grosse, G., Jones, B. M., and Anthony, K. W. (2012). Peat accumulation in drained thermokarst lake basins in continuous, ice-rich permafrost, northern Seward Peninsula,

- Alaska. *Journal of Geophysical Research: Biogeosciences*, 117(G2). <https://doi.org/10.1029/2011JG001766>
- Jorgenson, M. T., Harden, J., Kanevskiy, M., O'Donnell, J., Wickland, K., Ewing, S., Manies, K., Zhuang, Q., Shur, Y., Striegl, R., and Koch, J. (2013). Reorganization of vegetation, hydrology and soil carbon after permafrost degradation across heterogeneous boreal landscapes. *Environmental Research Letters*, 8(3), 035017. <http://dx.doi.org/10.1088/1748-9326/8/3/035017>
- Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., O'Donnell, J., Schuur, E. A. G., Kanevskiy, M., and Marchenko, S. (2010). Resilience and vulnerability of permafrost to climate change. *Canadian Journal of Forest Research*, 40(7), 1219-1236. <https://doi.org/10.1139/X10-060>
- Kasischke, E. S., Verbyla, D. L., Rupp, T. S., McGuire, A. D., Murphy, K. A., Jandt, R., Barnes, J. L., Hoy, E. E., Duffy, P. A., Calef, M., and Turetsky, M. R. (2010). Alaska's changing fire regime - implications for the vulnerability of its boreal forests. *Canadian Journal of Forest Research*, 40(7), 1313-1324. <https://doi.org/10.1139/X10-098>
- Kawahigashi, M., Prokushkin, A., and Sumida, H. (2011). Effect of fire on solute release from organic horizons under larch forest in Central Siberian permafrost terrain. *Geoderma*, 166(1), 171-180. <https://doi.org/10.1016/j.geoderma.2011.07.027>
- Kettridge, N., Humphrey, R. E., Smith, J. E., Lukenback, M. C., Devito, K. J., Petrone, R. M., and Waddington, J. M. (2014). Burned and unburned peat water repellency: implications for peatland evapotranspiration following wildfire. *Journal of Hydrology*, 513, 335-341. <https://doi.org/10.1016/j.jhydrol.2014.03.019>
- Koch, J. C., Kikuchi, C. P., Wickland, K. P., and Schuster, P. (2014). Runoff sources and flow paths in a partially burned, upland boreal catchment underlain by permafrost. *Water Resour Res*, 50(10), 8141-8158. <https://doi.org/10.1002/2014WR015586>
- Kong, J., and Yang, J. (2014). Short-and long-term effects of fire on soil properties in a Dahurian larch forest in Great Xing'an Mountains. *Chinese Journal of Ecology*, 33(6), 1445-1450 (in Chinese).
- Kopp, B. J., Lange, J., and Menzel, L. (2017). Effects of wildfire on runoff generating processes in northern Mongolia. *Regional Environmental Change*, 17, 1951-1963. <https://doi.org/10.1007/s10113-016-0962-y>
- Köster, K., Berninger, F., Heinonsalo, J., Lindén, A., Köster, E., Ilvesniemi, H., and Pumpanen, J. (2015). The long-term impact of low-intensity surface fires on litter decomposition and enzyme activities in boreal coniferous forests. *International Journal of Wildland Fire*, 25(2), 213-223. <https://doi.org/10.1071/WF14217>

- Kwok, R. (2018). Arctic sea ice thickness, volume, and multiyear ice coverage: losses and coupled variability (1958–2018). *Environmental Research Letters*, 13(10), 105005. <https://doi.org/10.1088/1748-9326/aae3ec>
- Lewis, T. L., Lindberg, M. S., Schmutz, J. A., and Bertram, M. R. (2014). Multi-trophic resilience of boreal lake ecosystems to forest fires. *Ecology*, 95(5), 1253-1263. <https://doi.org/10.1890/13-1170.1>
- Lewkowicz, A. G., Etzelmüller, B., and Smith, S. L. (2011). Characteristics of discontinuous permafrost based on ground temperature measurements and electrical resistivity tomography, southern Yukon, Canada. *Permafrost and Periglacial Processes*, 22(4), 320-342. <https://doi.org/10.1002/ppp.703>
- Lewkowicz, A. G., and Harris, C. (2005). Morphology and geotechnique of active-layer detachment failures in discontinuous and continuous permafrost. *Geomorphology*, 69(1-4), 275-297. <https://doi.org/10.1016/j.geomorph.2005.01.011>
- Li, X., Jin, H., He, R., Huang, Y., Wang, H., Luo, D., Jin, X., Lü, L., Wang, L., Li, W., Wei, C., Chang, X., Yang, S., and Yu, S. (2019). Effects of forest fires on the permafrost environment in the northern Da Xing'anling (Hinggan) mountains, Northeast China. *Permafrost and Periglacial Processes*, 30(3), 163-177. <https://doi.org/10.1002/ppp.2001>
- Lipovsky, P. S., Coates, J., Lewkowicz, A. G., and Trochim, E. (2005). Active layer detachments following the summer 2004 forest fires near Dawson City, Yukon. In D. S. Edmond, G. D. Bradshaw, L. L. Lewis, and L. H. Weston (Eds.), *Yukon Exploration and Geology* (pp. 175-194). Yukon Geological Survey.
- Loranty, M. M., Abbott, B. W., Blok, D., Douglas, T. A., Epstein, H. E., Forbes, B. C., Jones, B. M., Kholodov, A. L., Kropp, H., Malhotra, A., Mamet, S. D., Myers-Smith, I. H., Natali, S. M., O'Donnell, J. A., Phoenix, G. K., Rocha, A. V., Sonnentag, O., Tape, K. D., and Walker, A. D. (2018). Reviews and syntheses: Changing ecosystem influences on soil thermal regimes in northern high-latitude permafrost regions. *Biogeosciences*, 15(17), 5287-5313. <https://doi.org/10.5194/bg-15-5287-2018>
- Loranty, M. M., Natali, S. M., Berner, L. T., Goetz, S. J., Holmes, R. M., Davydov, S. P., Zimov, N. S., and Zimov, S. A. (2014). Siberian tundra ecosystem vegetation and carbon stocks four decades after wildfire. *Journal of Geophysical Research: Biogeosciences*, 119(11), 2144–2154. <https://doi.org/10.1002/2014JG002730>
- Mackay, J. R. (1995). Active layer changes (1968 to 1993) following the forest-tundra fire near Inuvik, N.W.T., Canada. *Arctic and Alpine Research*, 27(4), 323-336.
- Manasypov, R. M., Shirokova, L. S., and Pokrovsky, O. S. (2017). Experimental modelling of thaw lake water evolution in discontinuous permafrost zone: Role of peat, lichen leaching and ground fire. *Science of the Total Environment*, 580, 245-257. <https://doi.org/10.1016/j.scitotenv.2016.12.067>

- McGuire, A. D., Macdonald, R. W., Schuur, E. A. G., Harden, J. W., Kuhry, P., Hayes, D. J., Christensen, T. R., and Heimann, M. (2010). The carbon budget of the northern cryosphere region. *Current Opinion in Environmental Sustainability*, 2(4), 231-236. <https://doi.org/10.1016/j.cosust.2010.05.003>
- Mekonnen, Z. A., Riley, W. J., Randerson, J. T., Grant, R. F., and Rogers, B. M. (2019). Expansion of high-latitude deciduous forests driven by interactions between climate warming and fire. *Nature Plants*, 5, 952-958. <https://doi.org/10.1038/s41477-019-0495-8>
- Michaelides, R. J., Schaefer, K., Zebker, H. A., Parsekian, A., Liu, L., Chen, J., Natali, S., Ludwig, S., and Schaefer, S. R. (2019). Inference of the impact of wildfire on permafrost and active layer thickness in a discontinuous permafrost region using the remotely sensed active layer thickness (ReSALT) algorithm. *Environmental Research Letters*, 14(3), 035007. <https://doi.org/10.1088/1748-9326/aaf932>
- Minsley, B. J., Pastick, N. J., Wylie, B. K., Brown, D. R. N., and Kass, M. A. (2016). Evidence for nonuniform permafrost degradation after fire in boreal landscapes. *Journal of Geophysical Research: Earth Surface*, 121(2), 320-335. <https://doi.org/10.1002/2015JF003781>
- Myers-Smith, I. H., and Hik, D. S. (2018). Climate warming as a driver of tundra shrubline advance. *Journal of Ecology*, 106(2), 547-560. <https://doi.org/10.1111/1365-2745.12817>
- Neumann, R. B., Moorberg, C. J., Lundquist, J. D., Turner, J. C., Waldrop, M. P., McFarland, J. W., Euskirchen, E. S., Edgar, C. W., Turetsky, M. R. (2019). Warming effects of spring rainfall increase methane emissions from thawing permafrost. *Geophysical Research Letters*, 46(3), 1393-1401. <https://doi.org/10.1029/2018GL081274>
- Northern Climate Exchange. (2013). *Burwash Landing and Destruction Bay landscape hazards: geological mapping for climate change adaptation planning*. Yukon Research Centre, Yukon College. https://www.yukonu.ca/sites/default/files/inline-files/Hazards_BurwashDBay_finalApr29_web.pdf
- Nossov, D. R., Jorgenson, M. T., Kielland, K., and Kanevskiy, M. Z. (2013). Edaphic and microclimatic controls over permafrost response to fire in interior Alaska. *Environmental Research Letters*, 8(3), 035013. <https://doi.org/10.1088/1748-9326/8/3/035013>
- Obu, J., Westermann, S., Kääb, A., and Bartsch, A. (2018). *Ground Temperature Map, 2000-2016, Northern Hemisphere Permafrost*. PANGAEA, Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, Bremerhaven, Germany. <https://doi.org/10.1594/PANGAEA.888600>.
- O'Donnell, J. A., Harden, J. W., McGuire, A. D., Kanevskiys, M. Z., Jorgenson, M. T., and Xu, X. (2011a). The effect of fire and permafrost interactions on soil carbon accumulation in an upland black spruce ecosystem of interior Alaska: implications for post-thaw carbon

- loss. *Global Change Biology*, 17(3), 1461-1474. <https://doi.org/10.1111/j.1365-2486.2010.02358.x>
- O'Donnell, J. A., Harden, J. W., McGuire, A. D., and Romanovsky, V. E. (2011b). Exploring the sensitivity of soil carbon dynamics to climate change, fire disturbance and permafrost thaw in a black spruce ecosystem. *Biogeosciences*, 8, 1367-1382. <https://doi.org/10.5194/bg-8-1367-2011>
- O'Donnell, J. A., Jorgenson, M. T., Harden, J. W., McGuire, A. D., Kanevskiy, M. Z., and Wickland, K. P. (2012). The effects of permafrost thaw on soil hydrologic, thermal, and carbon dynamics in an Alaskan Peatland. *Ecosystems*, 15, 213-229. <https://doi.org/10.1007/s10021-011-9504-0>
- Olefeldt, D., Goswami, S., Grosse, G., Hayes, D., Hugelius, G., Kuhry, P., McGuire, A. D., Romanovsky, V. E., Sannel, A. B. K., Schuur, E. A. G., Turetsky, M. R. (2016). Circumpolar distribution and carbon storage of thermokarst landscapes. *Nature Communications*, 7, 13043. <https://doi.org/10.1038/ncomms13043>
- O'Neill, H. B., Roy-Léveillé, P., Lebedeva, L., and Ling, F. (2020). Recent advances (2010-2019) in the study of taliks. *Permafrost and Periglacial Processes*, 31(3), 346-357. <https://doi.org/10.1002/ppp.2050>
- Owens, P. N., Giles, T. R., Petticrew, E. L., Leggat, M. S., Moore, R. D., and Eaton, B. C. (2013). Muted responses of streamflow and suspended sediment flux in a wildfire-affected watershed. *Geomorphology*, 202, 128-139. <https://doi.org/10.1016/j.geomorph.2013.01.001>
- Pastick, N. J., Jorgenson, M. T., Wylie, B. K., Minsley, B. J., Ji, L., Walvoord, M. A., Smith, B. D., Abraham, J. D., and Rose, J. R. (2013). Extending airborne electromagnetic surveys for regional active layer and permafrost mapping with remote sensing and ancillary data, Yukon Flats Ecoregion, Central Alaska. *Permafrost and Periglacial Processes*, 24(3), 184-199. <https://doi.org/10.1002/ppp.1775>
- Rocha, A. V., Loranty, M. M., Higuera, P. E., Mack, M. C., Hu, F. S., Jones, B. M., Breen, A. L., Rastetter, E. B., Goetz, S. J., and Shaver, G. R. (2012). The footprint of Alaskan tundra fires during the past half-century: Implications for surface properties and radiative forcing. *Environmental Research Letters*, 7(4), 044039. <https://doi.org/10.1088/1748-9326/7/4/044039>
- Rocha, A. V., and Shaver, G. R. (2011). Postfire energy exchange in arctic tundra: the importance and climatic implications of burn severity. *Global Change Biology*, 17(9), 2831-2841. <https://doi.org/10.1111/j.1365-2486.2011.02441.x>
- Schuster, P. F., Striegl, R. G., Aiken, G. R., Krabbenhoft, D. P., Dewild, J. F., Butler, K., Kamark, B., and Dornblaser, M. (2011). Mercury export from the Yukon River Basin and

- potential response to a changing climate. *Environmental Science & Technology*, 45(21), 9262-9267. <https://doi.org/10.1021/es202068b>
- Seibert, J., McDonnell, J. J., and Woodsmith, R. D. (2010). Effects of wildfire on catchment runoff response: A modelling approach to detect changes in snow-dominated forested catchments. *Hydrology Research*, 41(5), 378-390. <https://doi.org/10.2166/nh.2010.036>
- Shur, Y. L., and Jorgenson, M. T. (2007). Patterns of permafrost formation and degradation in relation to climate and ecosystems. *Permafrost and Periglacial Processes*, 18(1), 7-19. <https://doi.org/10.1002/ppp.582>
- Smith, S. L., Riseborough, D. W., and Bonnaventure, P. P. (2015). Eighteen-year record of forest fire effects on ground thermal regimes and permafrost in the central Mackenzie Valley, NWT, Canada. *Permafrost and Periglacial Processes*, 26(4), 289-303. <https://doi.org/10.1002/ppp.1849>
- Sun, H., Santalahti, M., Pumpanen, J., Köster, K., Berninger, F., Raffaello, T., Jumpponen, A., Asiegbu, F. O., and Heinonsalo, J. (2015). Fungal community shifts in structure and function across a boreal forest fire chronosequence. *Applied and Environmental Microbiology*, 81, 7869-7880. <https://doi.org/10.1128/AEM.02063-15>
- Swanson, D. K. (1996). Susceptibility of permafrost soils to deep thaw after forest fires in interior Alaska, U.S.A., and some ecologic implications. *Arctic and Alpine Research*, 28(2), 217-227.
- Tank, S. E., Olefeldt, D., Quinton, W. L., Spence, C., Dion, N., Ackley, C., Burd, K., Hutchins, R., and Mengistu, S. (2018). Fire in the Arctic: The effect of wildfire across diverse aquatic ecosystems of the Northwest Territories. In *Aqhaliat* (pp. 31-38), Polar Knowledge Canada. <https://www.canada.ca/en/polar-knowledge/aqhaliat.html>
- Tank, S. E., Vonk, J. E., Walvoord, M. A., McClelland, J. W., Laurion, I., and Abbott, B. W. (2020). Landscape matters: Predicting the biogeochemical effects of permafrost thaw on aquatic networks with a state factor approach. *Permafrost and Periglacial Processes*, 31(3), 358-370. <https://doi.org/10.1002/ppp.2057>
- Taş, N., Prestat, E., McFarland, J. W., Wickland, K. P., Knight, R., Berhe, A. A., Jorgenson, T., Waldrop, M. P., and Jansson, J. K. (2014). Impact of fire on active layer and permafrost microbial communities and metagenomes in an upland Alaskan boreal forest. *The ISME Journal*, 8(9), 1904-1919. <https://doi.org/10.1038/ismej.2014.36>
- Turetsky, M. R., Donahue, W. F., and Benscoter, B. W. (2011). Experimental drying intensifies burning and carbon losses in a northern peatland. *Nature Communications*, 2, 514. <https://doi.org/10.1038/ncomms1523>
- Turetsky, M. R., Mack, M. C., Hollingsworth, T. N., and Harden, J. W. (2010). The role of mosses in ecosystem succession and function in Alaska's boreal forest. *Canadian Journal of Forest Research*, 40(7), 1237-1264. <https://doi.org/10.1139/X10-072>

- Veraverbeke, S., Rogers, B. M., Goulden, M. L., Jandt, R. R., Miller, C. E., Wiggins, E. B., and Randerson, J. T. (2017). Lightning as a major driver of recent large fire years in North American boreal forests. *Nature Climate Change*, 7, 529-534. <https://doi.org/10.1038/nclimate3329>
- Verma, S., and Jayakumar, S. (2012). Impact of forest fire on physical, chemical and biological properties of soil: A review. *Proceedings of the International Academy of Ecology and Environmental Sciences*, 2(3), 168-176.
- Walker, X. J., Mack, M. C., and Johnstone, J. F. (2015). Stable carbon isotope analysis reveals widespread drought stress in boreal black spruce forests. *Global Change Biology*, 21(8), 3102-3113. <https://doi.org/10.1111/gcb.12893>
- Walker, X. J., Baltzer, J. L., Cumming, S. G., Day, N. J., Johnstone, J. F., Rogers, B. M., Solvik, K., Turetsky, M. R., and Mack, M. C. (2018a). Soil organic layer combustion in boreal black spruce and jack pine stands of the Northwest Territories, Canada. *International Journal of Wildland Fire*, 27(2), 125-134. <https://doi.org/10.1071/WF17095>
- Walker, X. J., Rogers, B. M., Baltzer, J. L., Cumming, S. R., Day, N. J., Goetz, S., Johnstone, J. F., Schuur, E. A. G., Turetsky, M. R., and Mack, M. C. (2018b). Cross-scale controls on carbon emissions from boreal mega-fires of the Northwest Territories, Canada. *Global Change Biology*, 24(9), 4251-4265. <https://doi.org/10.1111/gcb.14287>
- Walker, X. J., Baltzer, J. L., Cumming, S. G., Day, N. J., Goetz, S., Johnstone, J. F., Potter, S., Rogers, B. M., Schuur, E. A. G., Turetsky, M. R., and Mack, M. C. (2019). Increasing wildfires threaten historic carbon sink of boreal forest soils. *Nature*, 572, 520-523. <https://doi.org/10.1038/s41586-019-1474-y>
- Walz, J., Knoblauch, C., Böhme, L., and Pfeiffer, E.-M. (2017). Regulation of soil organic matter decomposition in permafrost-affected Siberian tundra soils – Impact of oxygen availability, freezing and thawing, temperature, and labile organic matter. *Soil Biology and Biochemistry*, 110, 34-43. <https://doi.org/10.1016/j.soilbio.2017.03.001>
- Wang, X., Thompson, D. K., Marshall, G. A., Tymstra, C., Carr, R., and Flannigan, M. D. (2015). Increasing frequency of extreme fire weather in Canada with climate change. *Climatic Change*, 130, 573-586. <https://doi.org/10.1007/s10584-015-1375-5>
- Wotton, B. M., Flannigan, M. D., and Marshall, G. A. (2017). Potential climate change impacts on fire intensity and key wildfire suppression thresholds in Canada. *Environmental Research Letters*, 12(9), 095003. <https://doi.org/10.1088/1748-9326/aa7e6e>
- Yi, S., McGuire, A. D., Kasischke, E., Harden, J., Manies, K., Mack, M., and Turetsky, M. (2010). A dynamic organic soil biogeochemical model for simulating the effects of wildfire on soil environmental conditions and carbon dynamics of black spruce forests. *Journal of Geophysical Research: Biogeosciences*, 115(G4). <https://doi.org/10.1029/2010JG001302>

- Yoshikawa, K., Bolton, W. R., Romanovsky, V. E., Fukuda, M., and Hinzman, L. D. (2002). Impacts of wildfire on the permafrost in the boreal forests of Interior Alaska. *Journal of Geophysical Research: Atmospheres*, 107(D1), FFR 4-1-FFR 4-14. <https://doi.org/10.1029/2001JD000438>
- Zhang, Y., Wolfe, S. A., Morse, P. D., Olthof, I., and Fraser, R. H. (2015). Spatiotemporal impacts of wildfire and climate warming on permafrost across a subarctic region, Canada. *Journal of Geophysical Research: Earth Surface*, 120(11), 2338-2356. <https://doi.org/10.1002/2015JF003679>
- Zhou, X., Sun, H., Pumpanen, J., Sietiö, O.-M., Heinonsalo, J., Köster, K., and Berninger, F. (2019). The impact of wildfire on microbial C: N: P stoichiometry and the fungal-to-bacterial ratio in permafrost soil. *Biogeochemistry*, 142, 1-17. <https://doi.org/10.1007/s10533-018-0510-6>
- Zoltai, S. C. (1993). Cyclic development of permafrost in the peatlands of northwestern Alberta, Canada. *Arctic and Alpine Research*, 25(3), 240-246.

Note: the following references were not cited in the published paper: Swanson, 1996; Zoltai, 1993.

CHAPTER 3: HALF A CENTURY OF DISCONTINUOUS PERMAFROST PERSISTENCE AND DEGRADATION IN WESTERN CANADA

Abstract: Long-term field studies of permafrost change are needed to validate predictive models but few are possible because of a paucity of direct observations prior to the late 1970s. To help fill this knowledge gap, we resurveyed a transect of 68 sites, originally investigated in 1962, to evaluate change in the isolated patches and sporadic discontinuous permafrost zones between Keg River, Alberta (57.8°N) and Hay River, Northwest Territories (60.8°N). The goal was to establish the degree of permafrost degradation due to approximately 2°C of regional climate warming over the intervening 55 years, compounded at some sites by forest fire. By 2017-2018, permafrost had degraded at 36% of the 44 sites which exhibited it in 1962, but had persisted at a minimum of 50% with a further 14% potentially retaining permafrost. This is much less degradation than reported for a 1988-1989 survey of the same transect. Permafrost was maintained under thicker organic layers (86% > 40 cm) and at the majority of sites with fine-grained substrates, while degradation occurred preferentially at sites with coarse soils and thinner organic layers. Forest fire did not enhance the degree of permafrost loss, but greater frost table depths were observed at some burned locations. This study demonstrates that while the trajectory of change is towards permafrost loss, thin and warm permafrost can be persistent, even when disturbed. It also underlines the importance of considering the range of landscape types when projecting the rate of future permafrost thaw.

3.1 Introduction

Global air temperatures are expected to rise by 3.8 to 6°C by 2100 (under RCP 8.5) and precipitation, either as rainfall or snow cover, is also expected to increase (AMAP, 2017). These global changes are expected to be amplified in northern latitudes and to severely impact permafrost. For Canada, models suggest that by the end of the 21st century active layer thickness will increase, supra-permafrost taliks will develop, and permafrost may thaw entirely in southern regions, where it is warmer than -1°C and less than 5 m thick (Bush and Lemmen, 2019; Smith et al., 2004; Zhang et al., 2008). These changes in turn will impact the terrain (Lantz and Kokelj, 2008; Jorgenson et al., 2010), basin hydrology (Connon et al., 2015; Quinton et al., 2009; Woo, 1990), northern ecosystems (Baltzer et al., 2014; Jorgenson et al., 2001; Quinton et al., 2001), and infrastructure (Doré et al., 2016; Nelson et al., 2002), and will feedback to climate through

emissions of greenhouse gases from thawing permafrost (McGuire et al., 2010; O'Donnell et al., 2011; Schuur et al., 2008; Turetsky et al., 2002).

Several studies demonstrate that changes to permafrost have occurred over the last three decades (Gibson et al., 2018; Quinton et al., 2001; Robinson et al., 2000; Romanovsky et al., 2010; Smith et al., 2010). Evaluations of change in permafrost distribution over longer periods in North America are very limited, however, because few field data were collected prior to the 1970s, especially in relation to permafrost distribution near its southernmost limit. Such comparisons (e.g. James et al., 2013; Kwong and Gan, 1994) are needed to support modelled predictions of permafrost response to future climate change (AMAP, 2017).

R.J.E. Brown conducted an extensive survey of permafrost along the Mackenzie Highway in northern Alberta and southern Northwest Territories (NWT), Canada in 1962 (Brown, 1964). Permafrost in this region extends from its southernmost margin, through the isolated patches zone (<10% permafrost by area) and across most of the sporadic discontinuous zone (10-50%) (Heginbottom et al., 1995). Permafrost in these zones is thin and close to 0°C and is therefore regarded as sensitive to environmental change (Smith et al., 2010). The regional climate began to warm in the mid-1960s (Kwong and Gan, 1994; Vincent et al., 2015) and forest fire, which represents an additional disturbance at some locations (Gibson et al., 2018), is becoming more frequent and severe as the climate warms (Wotton et al., 2010).

We re-examined the sites investigated by Brown in 1962 with the goal of evaluating current permafrost conditions, comparing loss rates to those observed elsewhere (e.g. James et al., 2013; Quinton et al., 2001; Smith et al., 2010), and assessing the role of forest fire in permafrost change. In so doing, we followed up on a survey conducted in 1988-1989 by Kwong and Gan (1994) who reported that permafrost along the same transect had entirely degraded south of Meander River, Alberta (59.05°N) and was present farther north at only 25% of the sites where Brown had observed it in 1962. They also projected significant additional permafrost degradation as the climate warmed further, as it has done over the last three decades.

3.2 Study area

The study area is a transect along the Mackenzie Highway and two adjacent spur roads, extending from near Keg River, Alberta (57.8°N) to Hay River, NWT (60.8°N) (Figure 3-1). This area falls in the taiga plains ecozone, and comprises flat to gently undulating terrain draped

in glaciolacustrine deposits and till. The vegetation is dominantly black spruce (*Picea mariana*) forest in the lowlands and mixed-wood forest in the uplands, and the understorey is commonly made up of mosses, lichens, sedges, and Labrador tea (*Rhododendron groenlandicum*) (Ecological Stratification Working Group, 1996). Wetlands make up approximately 40% of the area, comprising peat plateaus and thermokarst bogs, with peat depths between 2-6 m (Gibson et al., 2018). Peat plateaus are elevated 1-3 m above the surrounding bogs, with vegetation dominated by black spruce (*Picea mariana*), Labrador tea (*Rhododendron groenlandicum*), lichens, and *Sphagnum* (Gibson et al., 2018).

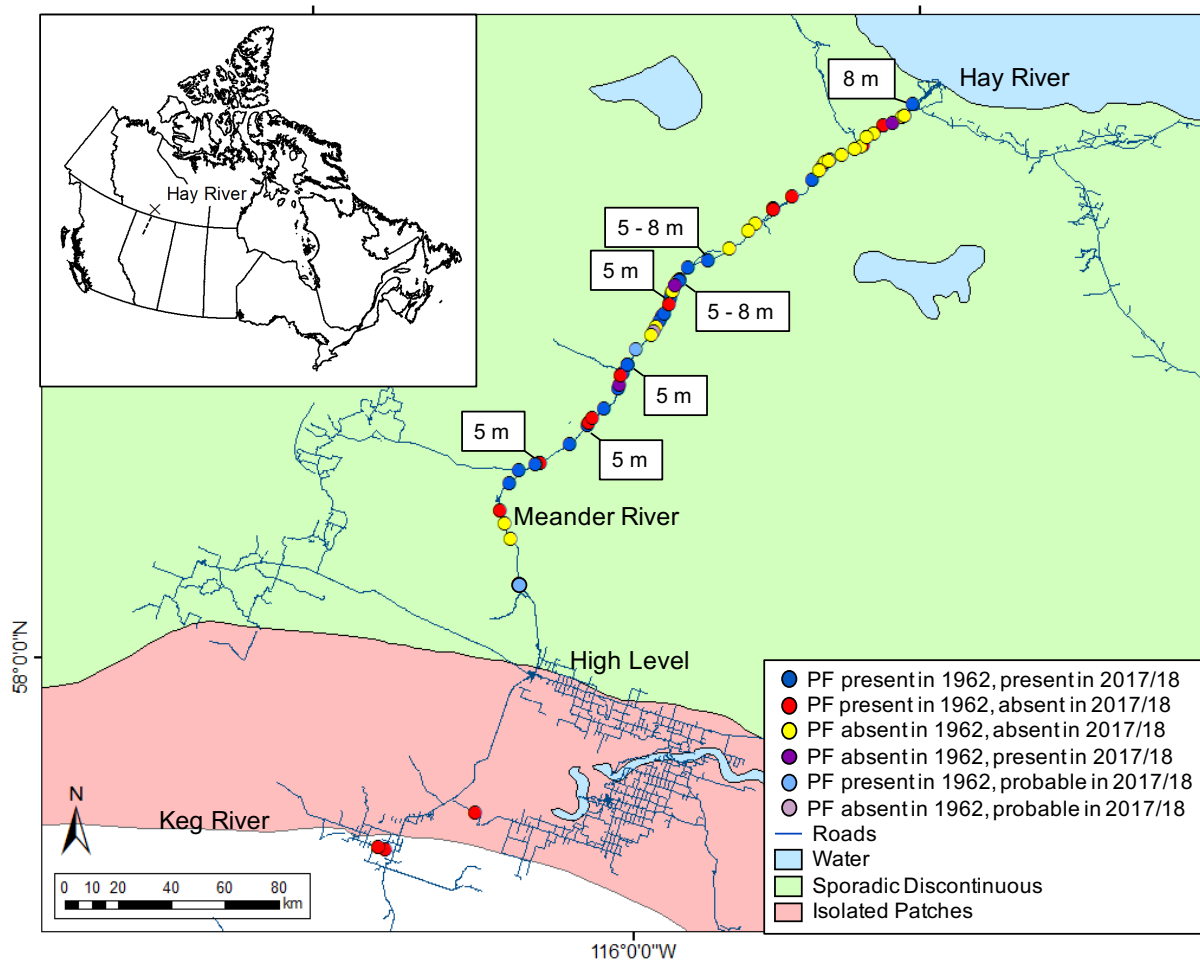


Figure 3-1: Study transect along the Mackenzie Highway showing changes in permafrost (PF) conditions between 1962 and 2017-2018 and permafrost zonation (Brown, 1964). Inferred permafrost thickness in 2017-2018 is shown at each site where an ERT survey was undertaken. Inset map shows location of the transect (dashed line) and Hay River, NT.

Permafrost in the sporadic discontinuous and isolated patches zones ranges from less than 15 m to greater than 60 m thick, with mean annual ground temperatures above -2.5°C (Smith et al., 2010). Permafrost occurs mostly in peatlands at the southern margin of this ecozone, and has been degrading over the last 100 years (Beilman and Robinson, 2003; Smith et al., 2005). Ground warming has been observed to the north of the study transect near Fort Simpson and Wrigley in recent decades (Smith et al., 2005) but there are no temperature monitoring boreholes along the transect itself.

The climate of the region is subarctic and continental, with mean annual air temperature (MAAT) for 1981-2010 of -2.5°C at Hay River and -1.0°C at High Level (Figure 3-2) (Environment Canada, 2018). Annual precipitation from 1981-2010 at Hay River is 336 mm and at High Level is 372 mm, about 40% of which falls as snow (Environment Canada, 2018). Brown's 1962 survey was preceded by approximately 15 years of relatively stable MAATs (Figure 3-2), but since then, the transect has experienced warming of about 2°C .

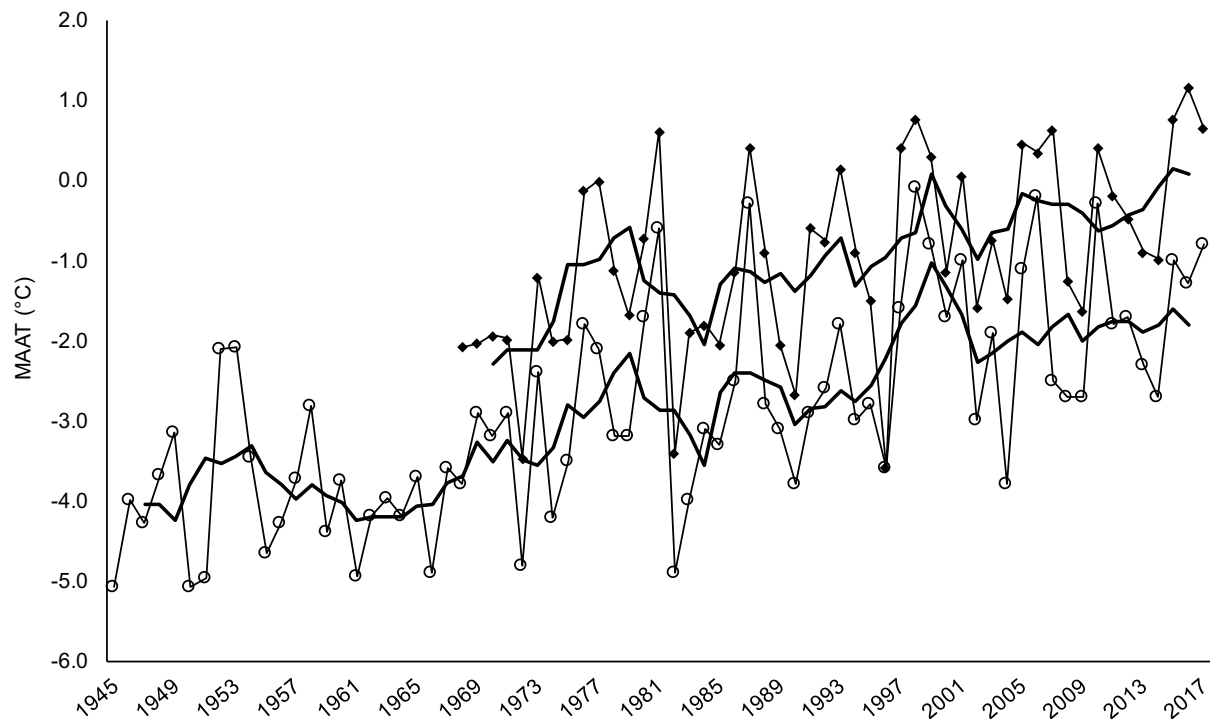


Figure 3-2: Mean annual air temperature (MAAT) at Hay River, NT (circles) from 1945 to 2017 and High Level, AB (diamonds) from 1967 to 2017 (data source: Environment Canada, 2018). The thick solid lines are 5-year running means.

3.3 Methods

Our goal was to evaluate permafrost presence and characteristics by revisiting the sites examined by Brown in September 1962 (Brown, 1964). In Brown (1964), these are shown on a map at a scale of 1:400,000 with site characteristics (vegetation, soil type, and relief) described in tabular form. Drainage conditions are mentioned only for a small number of sites. The map was digitized and overlaid on a Google Earth image to generate approximate site coordinates. Brown lists a total of 88 sites but indicates permafrost presence or absence for only 73. We were able to access 68 (93%) of these in August 2017 and 2018, with the remaining five omitted because of land use change to built-up area or farmland (three sites) or because access was impossible due to intervening surface water (two sites) (Table B1).

In the field, 2 km sweeps of the highway were conducted, centred on the approximate coordinates, and locations within the site area were identified as the best matches for Brown's descriptions of terrain, vegetation, and drainage (where given). These locations were probed using a 110 cm long frost probe, starting at the approximate coordinates in undisturbed terrain at least 50-100 m from the highway clearance. If a frost table was identified, depth to frozen ground was measured at 10 points located 1 m apart. As in the original survey (Brown, 1964), the observation of a frost table in late summer was considered to indicate that permafrost was present. If no frost table was reached, investigations continued at one to four other locations within the site area, identified by vegetation and drainage as potentially having permafrost, until a frost table was located or it was concluded that no frozen ground was present. Organic mat thickness and substrate were recorded at all transects, and soil samples were collected from the active layer at permafrost sites.

At sites where no frost table was identified, ground temperature profiles were recorded continuously for 15 minutes with an Onset Hobo UX120-006M Analog Data Logger (accuracy $\pm 0.15^{\circ}\text{C}$) linked to four thermistors attached at 25 cm increments on a wooden dowel that had been inserted into the ground to a minimum depth of 80 cm. If a measurement of 0°C or lower was reached after 15 minutes, permafrost was recorded as present. If not, the 15-minute values were used to estimate whether the frost table was present within 2 m of the ground surface by extrapolating the temperature gradient using linear regression (e.g. James et al., 2013), excluding measurements taken within 20 cm of the ground surface. If a frost table was predicted within 2 m but no physical observation of a frost table was made, the site was designated as being

permafrost probable (Table B2). The target depth of 2 m was chosen because the maximum depth that refreezes in this area is <2 m (Connon et al., 2018), 89% of frost table depths recorded by Brown were less than 1 m and except in river channels, the maximum frost table depth he recorded was 1.5 m (Brown, 1964). If no frost table was directly observed nor predicted using the temperature gradient, permafrost was deemed to be absent. This allows for direct comparison with the original survey which used similar methods and the same terminology (Brown, 1964). We recognize, however, that the term “permafrost absent” applies only to the top 2 m and that degrading permafrost could still be present at greater depths.

The outcome of these procedures was assignment of each site to one of four change / no change categories, two of which had sub-categories of present and probable: (1) permafrost present in 1962 and present or probable in 2017-2018; (2) permafrost present in 1962 and absent in 2017-2018; (3) permafrost absent in 1962 and in 2017-2018; (4) permafrost absent in 1962 and present or probable in 2017-2018.

Frost table depths were measured once for each site between August 6-16, 2017 or August 15-September 2, 2018. Since maximum thaw depth in the area occurs several weeks later based on ground temperature measurements on the west side of Great Slave Lake (Figure C2), these measurements represent a minimum active layer thickness. Brown conducted his survey between September 12-20, 1962. In order to compare active layer change, the 2017-2018 measurements therefore were adjusted by 2-20% according to the square root of additional air thawing degree days between the day of measurement and the date in 1962. No frost table was encountered within 110 cm at many probing spots which could mean that permafrost was absent or that the frost table depth exceeded 110 cm. Because of this uncertainty, we present median thaw depths rather than means, and include values greater than 110 cm in the sample.

Electrical resistivity tomography (ERT) surveys, 80-160 m long, were undertaken to evaluate permafrost thickness at a subset of sites spanning the transect and with detectable frost tables. The unique electrical properties of frozen ground make ERT a valuable tool for permafrost characterization, and the technique has been widely employed as a means to delimit permafrost (e.g. Fortier et al., 2008; Hilbich et al., 2009; Lewkowicz et al., 2011). An ABEM LS multi-electrode, two-dimensional resistivity system was used for the surveys. The electrodes were arranged in a Wenner array with a minimum spacing of 2 m, giving a depth of investigation of 12 m (80 m long surveys) or 24 m (160 m long surveys). Tomograms were produced using the

software package RES2DINV, which performs smoothness-constrained inversions using finite difference forward modelling (Loke, 1999). The robust inversion method was used as it is the most appropriate to portray the rapid changes in resistivity that occur at the boundaries between frozen and unfrozen soils. Iteration was carried out until the root-mean-square (RMS) error fell below 5%, or to the fifth iteration, whichever occurred first. Model sensitivity was calculated in RES2DINV (Loke et al., 2003), and model blocks with sensitivity less than 0.1 were excluded (Way et al., 2018). All profiles were topographically corrected before inversion using elevations obtained with a ZipLevel Pro-2000 altimeter (2 mm resolution).

To determine whether sites had burned in the last 55 years, we used fire history databases obtained from the Alberta Government and the Northwest Territories Centre for Geomatics (Alberta Government, 2018; GNWT, 2018). Fire history observations span from 1931 to 2017 in Alberta and from 1965 to 2016 in the NWT. Fire scar polygons were compared with site locations as well as site descriptions of fire scars to determine if the sites had burned. Although studies show that large parts of mapped fire areas may remain unburned (Fraser et al., 2004; Kasischke et al., 2010), for this analysis we assumed the entire area within the fire polygon was burned.

3.4 Results

Results focus on the changes in climate and permafrost conditions between 1962 and 2017-2018. The latter include permafrost presence or absence as well as frost table depths, comparison of sites which burned and which did not, and identifying differences in site characteristics where permafrost has persisted or degraded.

3.4.1 Climate changes

Trends in MAAT, average warm half-year (May-October) and cold half-year (November-April) temperatures, annual precipitation, and annual snowfall were examined for Hay River (1945-2017) and High Level (1968-2017), respectively located at the northern end and near the southern end of the transect (Figure 3-1). Periods before and after the 1962 (Brown, 1964) and 1988-1989 (Kwong and Gan, 1994) surveys were also examined: 1945-1962, 1962-1989, and 1989-2017 for Hay River; 1968-1989, and 1989-2017 for High Level. Given time lags between climate and permafrost response, antecedent climate trends give an indication of whether

changes in frozen ground conditions might have been underway at the time of the previous surveys in 1962 and 1988-1989, while the trends leading up to 2017 indicate if change can be expected to continue into the near future.

MAAT increased by 0.4°C/decade at both Hay River (1945-2017) and High Level (1968-2017) (Figure 3-2, Table 3-1). The 1962 survey was preceded by a period with no significant trend in air temperature at Hay River, but following the survey, there was a period of significant warming until 2017, including prior to and after the 1988-1989 survey. At High Level, there was no significant trend between 1968 and 1989, but significant warming took place following 1989 through to 2017. Similar temporal trends are evident for warm half-year and cold half-year air temperatures but for some periods, the warm half-year trend was significant while the cold half-year one was not, or vice versa (Table 3-1).

There was no significant trend in annual precipitation or annual snowfall over the entire record for either station (Table 3-1). However, total annual precipitation and total snowfall increased at Hay River between 1945-1962, while from 1962-2017 they showed no significant trend. Prior to the 1989 survey, Hay River showed no significant trend in annual precipitation but a significant decrease in total snowfall, while High Level showed no significant trend in either total precipitation or snowfall. Between 1989-2017, Hay River and High Level both showed no significant trend in total annual precipitation but a significant increase in total snowfall.

Table 3-1: Linear trends in climate variables over various periods for Hay River, NT and High Level, AB. Linear trends shown are significant at a 0.05 level or better. Data source: Environment Canada, 2018.

Location	Period of record	MAAT (°C)			Average warm half-year air temperature (°C)			Average cold half-year air temperature (°C)			Annual precipitation (mm)			Annual snowfall (cm)		
		β	R^2	p	β	R^2	p	β	R^2	p	β	R^2	p	β	R^2	p
Hay River	1945-2017	0.04	0.43	3.57e ⁻¹⁰	0.02	0.16	3.65e ⁻⁴	0.06	0.33	8.61e ⁻⁸	Not significant			Not significant		
	1945-1962	Not significant			Not significant			Not significant			11.20	0.59	1.96e ⁻⁴	9.12	0.61	1.28e ⁻⁴
	1962-2017	0.05	0.36	1.15e ⁻⁶	0.02	0.09	0.02	0.07	0.28	2.76e ⁻⁵	Not significant			Not significant		
	1962-1989	0.07	0.22	0.01	Not significant			0.12	0.26	0.006	Not significant			-3.2	0.30	0.003
	1989-2017	0.05	0.16	0.03	0.06	0.24	0.007	Not significant			Not significant			2.97	0.29	0.003
High Level	1968-2017	0.04	0.26	1.4e ⁻⁴	0.02	0.14	0.007	0.05	0.14	0.009	Not significant			Not significant		
	1968-1989	Not significant			Not significant			Not significant			Not significant			Not significant		
	1989-2017	0.06	0.19	0.018	0.05	0.25	0.005	Not significant			Not significant			1.88	0.11	0.08

3.4.2 Changes in near-surface permafrost distribution

Brown reported permafrost to be present at 63% of the 73 sites where he provided information on permafrost presence or absence (Brown, 1964). Of the 68 sites we visited, 65% exhibited permafrost in 1962 (Table 3-2). The subset of sites visited (93% of Brown's sites), therefore is considered to be representative of the original survey, although given the percentage of permafrost in the region (Heginbottom et al., 1995), both surveys are biased towards permafrost presence.

Figure 3-3 shows the proportion of the 68 sites in each of the four change / no change categories. In 1962, permafrost was present at 44 of the 68 sites (Brown, 1964), but by 2017-2018 it had degraded at 16 of these (36%) and persisted at 22 (50%) (Table 3-2, Table B1). Based on measurements of the active layer temperature gradient, permafrost probably persisted at an additional six sites (Table B2) bringing the potential total persistence rate within 2 m of the surface to 64%. In 1962 Brown found permafrost as far south as Keg River (57.79°N), whereas in 2017-2018 the southernmost confirmed occurrence of permafrost was just north of Meander River at 59.12°N.

Sites where permafrost had persisted since 1962 were all low relief areas and were dominated by black spruce (*Picea mariana*) stands, with some also having tamarack (*Larix laricina*) and jack pine (*Pinus banksiana*). These sites had understories comprised mainly of hummocky *Sphagnum* or feather moss, lichen, and Labrador tea (*Rhododendron groenlandicum*) forming a thick organic layer generally underlain by fine-grained soils (clay and silt) (Figure 3-4, Table B1). 86% of sites where permafrost persisted had an organic layer thickness greater than 40 cm, and all had organic layers greater than 24 cm (Figure 3-4a). Permafrost persisted from 1962 at 78% of sites with fine-grained substrates (Figure 3-4b) but degraded at the majority (56%) of sites with coarse-grained soils (sand and gravel) and thin organic layers (Figure 3-4, Table B1). Of the degraded sites, only 23% had organic layers thicker than 40 cm, and 31% had organic layers less than 20 cm. All three sites which had organic layers greater than 70 cm but where permafrost degraded (sites 32, 83, and 85) were on coarse substrates and, in the last two cases, were at the southernmost end of the transect where the climate is less favourable for permafrost.

Permafrost was present (three sites) or probable (one site) at a total of four sites which did not exhibit permafrost during Brown's survey (Figure 3-3). This was an unexpected result,

differing from the findings of a survey conducted at the same latitudes along the Alaska Highway in northern British Columbia and southern Yukon (James et al., 2013).

Table 3-2: Presence of near-surface permafrost in the study area. Number of sites is shown in parentheses.

Survey year	Number of sites	Permafrost in 1962		Permafrost in 2017-2018		
		Present	Absent	Present	Probable	Absent
1962	73	63% (46)	37% (27)	-	-	-
2017-2018	68	65% (44)	35% (24)	37% (25)	10% (7)	53% (36)

Note: Permafrost present and probable totals include respectively three sites and one site where permafrost was not observed in 1962.

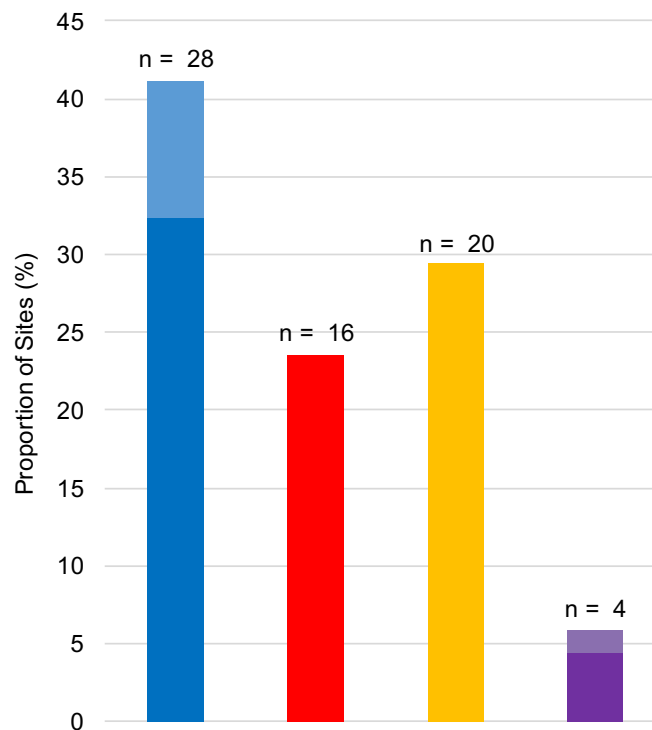


Figure 3-3: Proportion of the 68 sites in each change category: permafrost present in 1962 and present (blue) or probable (light blue) in 2017-2018; permafrost present in 1962 and absent in 2017-2018 (red); permafrost absent in 1962 and absent in 2017-2018 (yellow); permafrost absent in 1962 and present (purple) or probable (light purple) in 2017-2018.

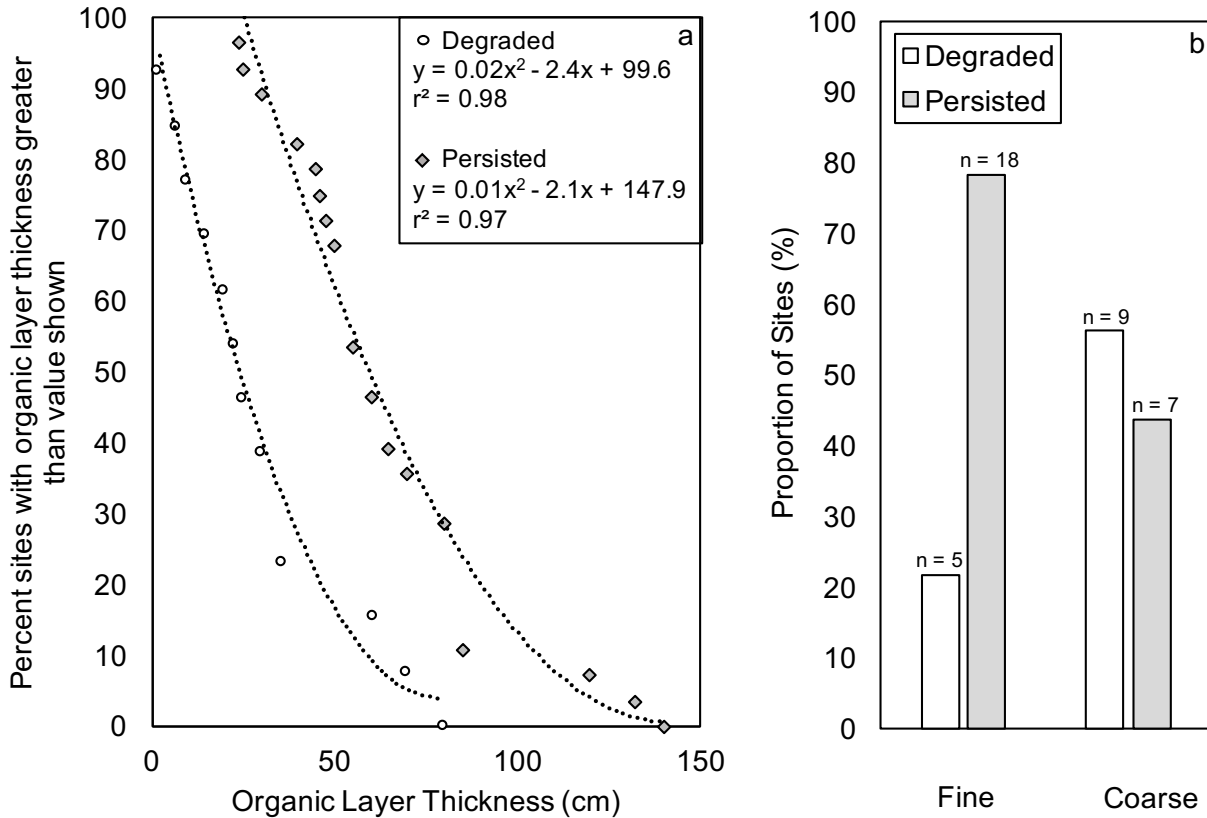


Figure 3-4: Relation between permafrost change (persistence or loss) from 1962 to 2017-2018 and (a) organic layer thickness and (b) substrate. Note: sites where permafrost was probable have been grouped with the sites where it persisted. No soil type information is available for five sites which had permafrost in 1962 as organic layers reached the frost table. The y-axis shows the percent of sites with organic layer thicknesses greater than the number shown on the x-axis. Organic layer thicknesses are minima because the frost table was shallower than the base of the organic layer at many permafrost sites and at some sites the entire depth of the organic layer was not excavated (maximum depth excavated was 140 cm).

3.4.3 Frost table depths

Frost table depths were highly variable in 2017-2018, both between and within sites (Figure 3-5). The mean frost table depth in 1962 was 64 cm, while for the same sites in 2017-2018 the median value was 66 cm. Intra-site variability in 2017-2018 ranged from as little as 11 cm to greater than 110 cm with a median value of 60 cm (Figure 3-5b). In the 1962 survey, intra-site variability ranged between 15 and 110 cm at five sites, and otherwise only a single value per site was reported (Brown, 1964). Many points are close to the 1:1 line in Figure 3-5a, indicating there has been little change in the population. Four sites show increases that exceed 20 cm (three of which have experienced fire) and two sites show apparent decreases of more than 20 cm. Sites where permafrost is probable show the greatest changes in frost table depths.

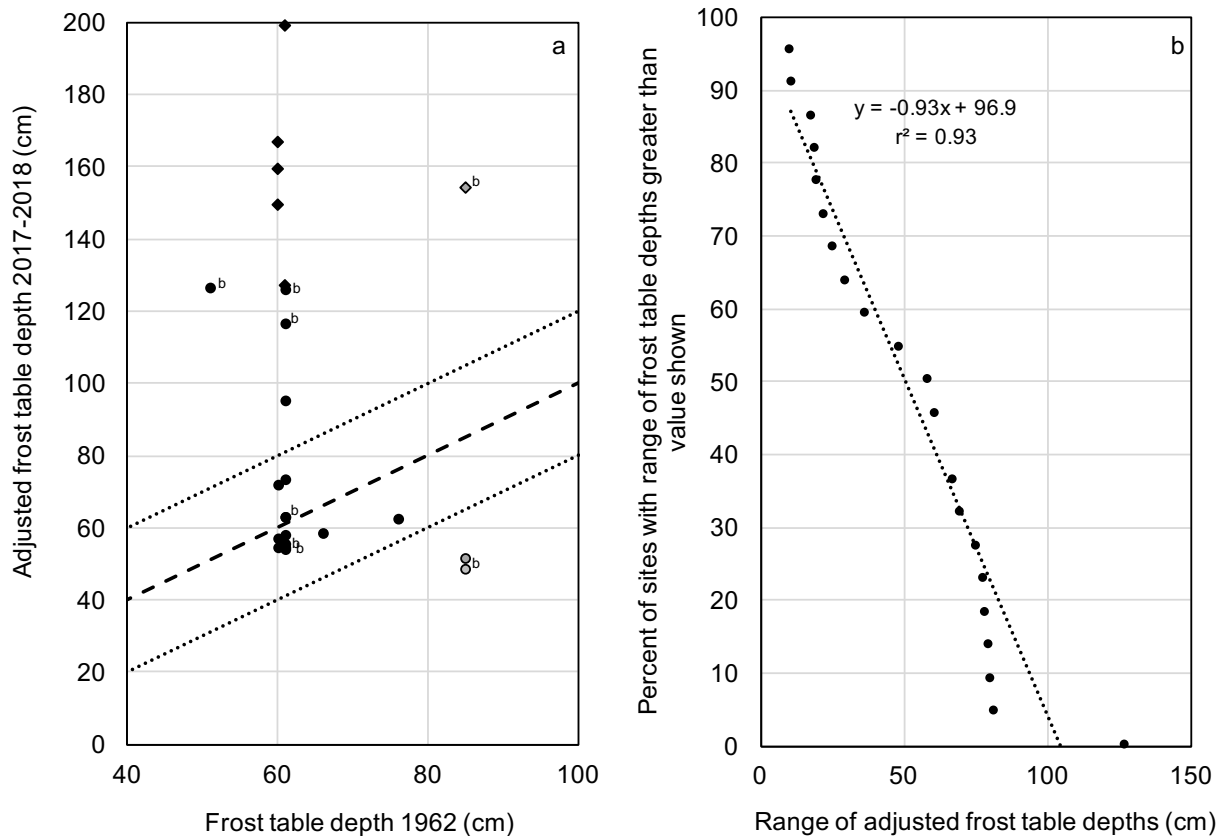


Figure 3-5: Frost table depths: (a) Comparison of site values for 1962 and 2017-2018 (Brown, 1964); (b) Intra-site ranges of adjusted frost table depths for 2017-2018. Note: the recent measurements have been adjusted according to the square root of additional thawing degree days between the day of the survey and the comparable date in 1962. In

(a), some 1962 values have been displaced by 1 cm to increase visibility. Dots represent the adjusted median values and diamonds represent the individual projected active layer depths at sites where permafrost was probable in 2017-2018. Grey symbols indicate sites where Brown (1964) provided a range of values (mean shown); all other sites were reported as individual measurements. The dashed line is the 1:1 line while the two dotted lines represent differences of ± 20 cm. Sites which burned between 1962 and 2017-2018 are indicated with a “b”.

Brown’s exact sampling locations and how many measurements per site was taken are not known, so inferences drawn from these results are tentative, but they suggest that there has been no major systematic change in active layer thickness since 1962.

3.4.4 Geophysical surveys

Electrical resistivity surveys were conducted at seven sites to validate permafrost presence and depth. Based on the resistivity measurements, permafrost was present in each case with a maximum thickness of 5-8 m (Figure 3-1, Table B1).

Figure 3-6 shows a typical ERT survey exhibiting several near-surface bodies of high resistivity that overlie a low resistivity layer extending to the base of the tomogram at 24 m.

Measurable (<110 cm) frost tables, where present, varied from 38-75 cm with a mean of 50 cm and a median of 49 cm (n=41), the latter increasing to 53 cm when samples > 110 cm are included (n = 61). There is good correspondence between measurable frost tables and the presence of higher resistivity bodies, indicating a boundary between frozen and unfrozen soil of approximately 350 Ωm , which gives a maximum permafrost thickness of approximately 5 m (Figure 3-6b). It is notable that the surface thawed layer at this site does not show up as a lower resistivity layer, presumably because it was too thin and/or too dry to greatly reduce the resistivity of the top layer of model blocks. It is also noteworthy that some locations where frozen ground is indicated by the ERT results did not have measurable frost tables, which supports our interpretation of the ongoing presence of permafrost beneath thicker active layers at the permafrost probable sites (Figure 3-5a; Table B1).

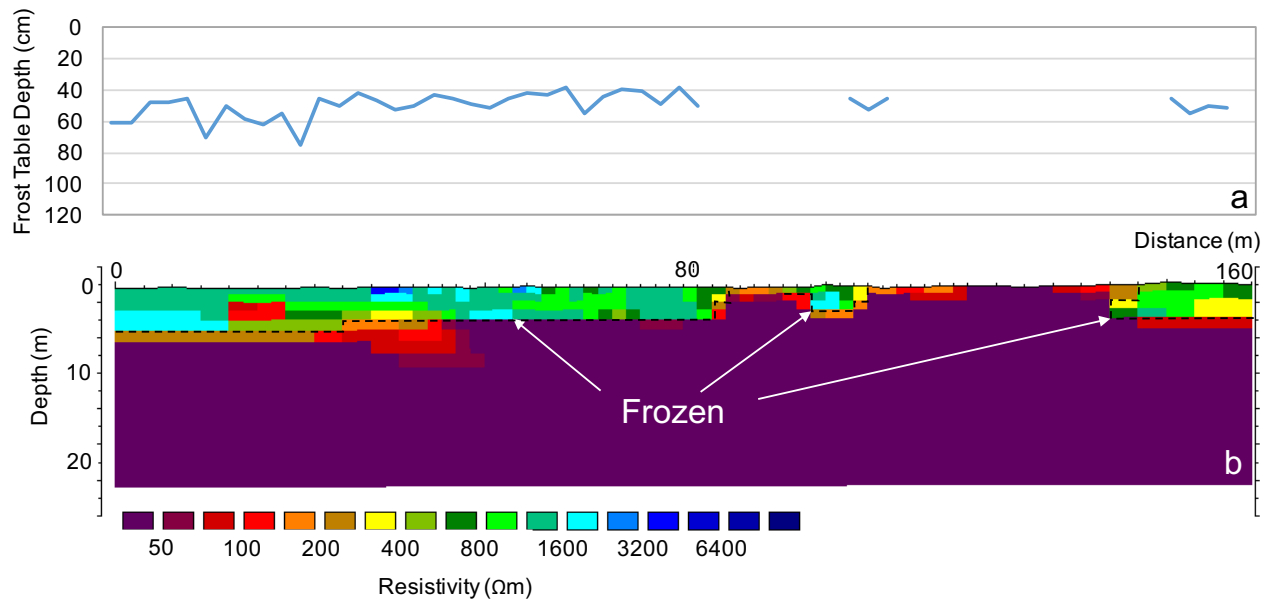


Figure 3-6: 160 m long permafrost survey undertaken at site 54 (59.9°N) on August 27, 2017: (a) Frost table depths measured at 2-4 m intervals (blank areas indicate no frost table within 110 cm of the surface), and; (b) ERT survey (5th iteration; RMS error 5.1%) shown as model blocks. Resistivity values >350 Ωm are interpreted as frozen ground which extends to a maximum depth of 6 m (dashed line).

3.4.5 Forest fire

Based on maps of fire scar polygons (Figure 3-7), 11 of the 68 sites (16%) burned during the last 55 years. The sites burned in five different years after the 1962 survey: 1971, 1981, 2000, 2007, and 2012. All but two of the sites which burned had permafrost present in 1962 (Table 3-3). Permafrost degraded at one burned site, persisted at seven, including site 68 which burned in both 2000 and 2012, with an additional site having probable permafrost. All of the sites where permafrost definitely persisted still have organic layers thicker than 50 cm.

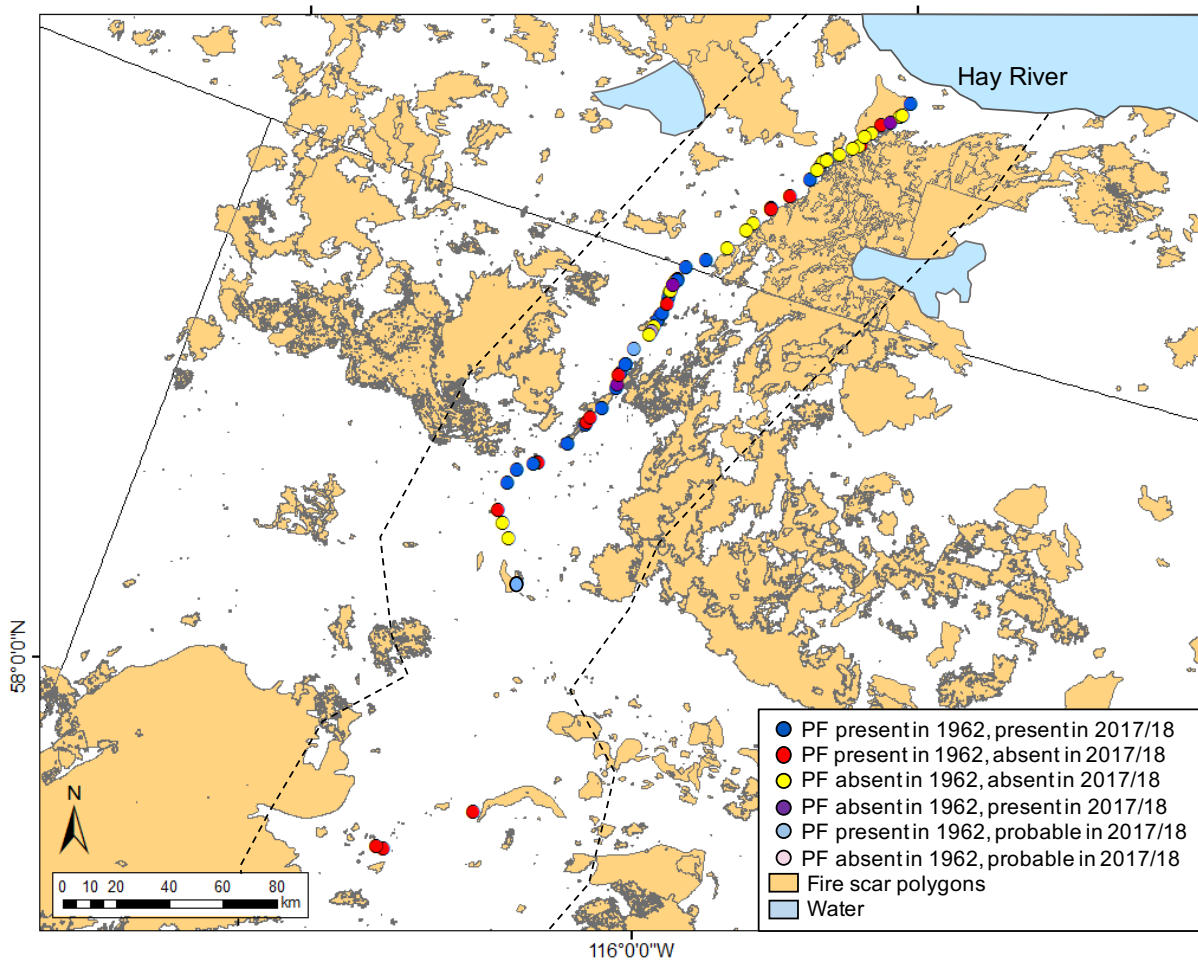


Figure 3-7: The study region and fire scar polygons from 1931-2017 for Alberta and 1965-2016 for NWT. Site locations and permafrost change between 1962 and 2017-2018 are indicated. (Data source: Alberta Government, 2018; GNWT, 2018). The 50 km radius around the transect used to calculate burned area is shown by dashed lines.

Table 3-3: Information about permafrost conditions and fire history at sites which burned between 1962 and 2017. (Data source: Alberta Government, 2018; GNWT, 2018). Note: dashes indicate sites for which no information is available for organic layer thickness in 1962.

Site	Latitude	Permafrost Present or Absent		Burn Year	Organic Layer Thickness (cm)	
		1962	2017-2018		1962	2017-2018
3	60.8	Present	Present	1971	91	> 85
5	60.7	Present	Probable	1971	-	25
6	60.7	Absent	Absent	1981	-	30
57	59.8	Present	Present	2012	> 61	> 65
58	59.8	Present	Present	2012	61	> 60
59	59.8	Absent	Absent	2012	-	1
68	59.5	Present	Present	2000 and 2012	112	> 50
69	59.5	Present	Present	2007	76	140
71	59.4	Present	Absent	2012	-	23
72	59.4	Present	Present	2012	> 61	> 50
73	59.3	Present	Present	2012	282	> 80

3.5 Discussion

3.5.1 Loss and persistence of permafrost

Our results show the potential for both permafrost persistence (50% of sites and probably 64%) and degradation (36% of sites, but possibly as high as 50%) along the transect between 1962 and 2017-2018. We favour the inclusion of the probable permafrost sites in the total for persistence, with the explanation that thick active layers prevented frost table detection by probing. Moreover, some sites classified as having lost permafrost in the top 2 m could have supra-permafrost taliks overlying frozen ground at greater depths. If this is the case, our totals over-represent the percentage permafrost loss, which further strengthens our argument that there is more permafrost remaining in the region than previously reported (Kwong and Gan, 1994).

We observed permafrost or considered it probable at four sites where it was not detected in 1962 (Table B1: sites 7, 50, 60 and 67). We propose two explanations for this unanticipated result. The first is that in these four instances we did not relocate the precise sites investigated by Brown. This explanation requires permafrost to have been present in the vicinity of the 1962

sites but not to have been observed. The second possibility is that permafrost aggraded in these locations between the survey dates, presumably because of vegetation succession or drainage change. We favour the first explanation given patchy permafrost distribution at the local scale (e.g. Figure 3-6) which makes it possible for frozen ground to have been missed by Brown, especially as aggradational features such as peat plateaus are absent at the four sites concerned. In contrast, the second possibility appears unlikely for several reasons. First, contemporary permafrost forms during late stages of ecological succession and generally only on river floodplains and drained-lake basins (Shur and Jorgenson, 2007). Three sites (50, 60 and 67) were already described as being in late successional stages when evaluated in 1962 with spruce canopies and well-developed organic layers so further changes resulting in permafrost aggradation appear unlikely. The fourth (site 7) was classified as a river or stream valley by Brown (1964), and when evaluated in 2018 it was a poorly drained bog with spruce trees and thick peat, so succession could have taken place there. Second, a change in drainage from wet to dry conditions which could also impact permafrost formation appeared to be possible at only one of the four locations (site 67), but vegetation remained similar and no aggradational permafrost features were observed. Based on the current climate, the significant warming over the record for this region (Figure 3-2, Table 3-2), and other evidence of regional permafrost degradation (Englefield, 1995; Robinson and Moore, 2000), we believe that the first explanation is the most likely to be correct.

The degree of permafrost change at sites north of Meander River was much less than the 75% loss reported from a widely cited survey carried out in 1988-1989 (Kwong and Gan, 1994). To the north of this location, permafrost persisted at 55% of the sites where Brown found it in 1962, with an additional 13% probably having permafrost. This results in a confirmed loss of only 32% of permafrost, less than half that previously reported for 1989. It appears highly unlikely that permafrost degraded at a large number of sites between 1962 and 1988-1989, and then aggraded at some of the same sites between 1989 and 2017-2018, especially considering the 1.4°C and 1.6°C increases in MAAT at Hay River and High Level respectively during the latter period (Figure 3-2). We therefore infer that the difference in findings is due to methodology. Kwong and Gan (1994) state that they attempted to examine as many of Brown's sites as possible using mileage markers as a guide, but they do not list the sites visited. They sampled 40 locations north of Meander River, representing 60% of Brown's sites and found permafrost at

only 10 sites. South of Meander River they examined 50% more sites than Brown, clearly showing that they did not limit their observations to Brown's original survey. Equally importantly, there is no indication that they checked additional locations in the vicinity of a prospective Brown site as we did if we could not locate a frost table on the first attempt (Kwong and Gan, 1994). Given inherent uncertainties in relocating Brown's sites because co-ordinates are not shown in his original report, and our much larger sample, we consider our results to be a more robust estimate of change from 1962. We are unable to conclusively concur with Kwong and Gan's (1994) statement that permafrost appears to be entirely absent south of Meander River because one of the sites we visited had probable permafrost (Figure 3-1).

Permafrost loss, where it has occurred, is attributed mainly to the 2°C climate warming that has taken place in the region since 1962 (Figure 3-2). Temperature increases have been two to three times as fast in the cold half-year (Table 3-1), but these may impact ground temperatures less than warm half-year increases due to buffering by the snowpack. There were no significant trends in annual precipitation over the complete record, but the increase in snowfall from 1989 to 2017-2018 at both Hay River and High Level may also have contributed to the observed permafrost changes.

There is no obvious spatial trend to permafrost loss and persistence within the sporadic discontinuous zone (Figure 3-1). The seemingly random pattern of degradation north of Meander River indicates that site-specific variables exert significant control on the timing of permafrost thaw. Based on the conceptual model of Shur and Jorgenson (2007), these sites fall into either ecosystem-driven or ecosystem-protected categories. Ecosystem-driven permafrost forms due to ecological succession as climate alone would not be sufficient for permafrost presence (Shur and Jorgenson, 2007). Ecosystem-protected permafrost formed under a past colder climate, but persists under late-successional ecosystems and cannot re-establish following disturbance (Shur and Jorgenson, 2007). MAATs for Hay River and High Level in the past decade are above -2°C, which indicates that most permafrost along the transect is probably ecosystem-protected.

The thin permafrost inferred from the ERT surveys suggests mean ground temperatures > -0.2°C at the depth of zero annual amplitude, and probably > -0.1°C. If these sites are in disequilibrium with the current climate, as seems likely given the ongoing warming (Figure 3-2), their thermal state is being maintained by the substantial heat inputs required to satisfy soil latent heat (French, 2018; Johnston et al., 1981; Kokelj and Burn, 2003). This explanation applies to

permafrost sites located north of the transect (Wolfe et al., 2015) and suggests that locations where permafrost has persisted to date would eventually thaw even if the climate did not warm further. With the degree of warming predicted, however, more rapid thaw of sites in the isolated patches and sporadic discontinuous zone can be expected (ACIA, 2005; Camil, 2005; Halsey et al., 1995) with permafrost persisting only at sites which are currently ecosystem-driven where MAAT remains below 2°C (Jorgenson et al., 2010; Shur and Jorgenson, 2007). Surprisingly, the latter may include some sites as far south as 59.3°N where permafrost has persisted after fire (Table 3-3).

At the site level, permafrost was observed in locations with black spruce (*Picea mariana*) canopies, thick organic layers (mostly >40 cm), fine-grained substrates and an absence of saturated conditions or surface ponding. These characteristics are typical of other locations where permafrost is present in the region (Connon et al., 2018; Gibson et al., 2018; Quinton et al., 2001) and in other similar areas (Nossov et al., 2013), and underline the importance of grain size, as well as the organic mat which has a low thermal conductivity when thawed and dry, and a high thermal conductivity when wet and frozen (Farouki, 1981; Williams and Smith, 1989). The result is little ground heat transfer in the summer months and enhanced heat loss during the winter, which favours the presence of permafrost.

The absence of a systematic change in frost table depths between 1962 and 2017-2018 (Figure 3-5a) suggests that permafrost is still fairly stable in the near surface at most sites, even though it could have warmed and thinned from the base upwards since 1962. However, at four of the five permafrost probable sites, the active layer is assessed as > 110 cm which is much thicker than observed by Brown for the same locations (Figure 3-5a). Degradation may therefore be under way at these locations.

Increased wetness and differences in drainage were observed qualitatively at some sites. Changes in drainage conditions can cause surface water to pool in depressions and enhance degradation, and groundwater carries heat to depth resulting in taliks (Connon et al., 2018; Woo, 1990). This can result in further degradation, surface subsidence, and additional water logging (Quinton et al., 2009; Woo, 1990). 56% of sites with coarse textures experienced permafrost loss, indicating the importance of substrate for trajectories of warm permafrost. Permafrost also degraded at some sites with thick organic layers, particularly at the southern end of the transect (Figure 3-1, Table 3-2), including at two sites with more than 80 cm of organics (Table B1).

Other field studies in this part of western Canada have found significant recent permafrost thaw (Connon et al., 2018; Gibson et al., 2018; Quinton et al., 2001). Quinton et al. (2001) reported that permafrost was present in approximately 70% of the Scotty Creek area (61.3°N) in 1947 and this decreased to approximately 43% by 2008. This represents degradation of almost 40% of the extant permafrost area in 1947. Connon et al. (2018) observed rapid permafrost degradation and talik formation at the same site. Our results are not directly comparable to these studies because (1) our sites extend across a 360 km long transect with heterogeneous terrain types and vegetation, (2) Brown's original sample was biased towards permafrost sites and (3) our data are point rather than areal in nature. Nevertheless, it is interesting to note that if the probable permafrost sites are considered to have permafrost present, the percentage of permafrost sites which degraded from 1962 to present is 36%, a value similar to the change at Scotty Creek. Unlike the latter, however, this permafrost loss has occurred at most sites without the development of obvious thermokarst. We attribute this to differences between the site-specific conditions of the Scotty Creek area compared to the Mackenzie Highway transect. The Scotty Creek area is composed of peatlands, partitioned into areas of permafrost peat plateaus, and non-permafrost channel fens, and flat bogs (Quinton et al., 2009), whereas Brown's sites are largely forested, only five were identified as peat plateaus and much drier conditions prevail than those reported for Scotty Creek.

3.5.2 Permafrost and forest fire

In total, 34% of the land area along the transect has been subject to forest fire since 1962 (Figure 3-7), so our sample of 11 out of 68 sites (16%) underrepresents the proportion of the landscape which has burned. However, fire does not appear to have had a notable impact in terms of overall permafrost presence or loss along this transect. Permafrost definitely degraded at site 71 which burned in 2012 (Table 3-3), and may have degraded at site 5 (listed as permafrost probable) which burned in 1971. Even if the latter is true, this is a lower rate of loss than for the transect as a whole. Organic layer thicknesses at sites 71 and 5 were 23 and 25 cm, respectively (Table 3-3). In contrast, organic layer thicknesses were all greater than 50 cm at sites where permafrost definitely persisted after fire. This result is similar to that of Jafarov et al. (2013) who found that for lowland black spruce sites near Fairbanks, Alaska, modelled permafrost degradation occurred only when the organic layer was less than 56 cm thick. The type of organic

material also affects permafrost persistence. Following fire, feathermoss has a higher thermal conductivity than *Sphagnum* because the latter's conductivity does not change significantly after burning (Yoshikawa et al., 2002). Most of the sites which burned and retained permafrost have a *Sphagnum* moss cover (Table B1). Site 3 is the exception to this, but it has > 85 cm of organic material and this proved sufficient to maintain permafrost regardless of organic material type. These results show that fire and climate warming do not necessarily result in permafrost degradation and that site-specific conditions and fire severity play important roles, especially in relation to the thickness of the residual organic layer (Camill and Clark, 1998; Halsey et al., 1995; Shur and Jorgenson, 2007; Smith et al., 2015).

Sites which burned exhibited thicker active layers in most cases, particularly those at the southern end of the transect which burned in 2012 (Figure 3-5, Table B1), suggesting that the more recent fires may have had an impact on this parameter. This supports Zhang et al. (2015) who found that active layer thickening following fire significantly increased after 1990 due to climate warming.

Changes in permafrost temperatures and active layer thickness as a result of fire have been well documented (Burn, 1998; Camill and Clark, 1998; Dyrness, 1982; Kasischke and Stocks, 2000; Mackay, 1995; Rouse, 1976; Viereck, 1982; Yoshikawa et al., 2002; Zhang et al., 2015;), but only recently for our study region (Fisher et al., 2016; Gibson et al., 2018). Thaw of permafrost in burned areas can be expected to continue until the insulating organic mat is re-established, and as the forest recovers, the surface energy balance and active layer thickness will return to pre-fire values (Camill and Clark, 1998; Dyrness, 1982; Jorgenson et al., 2010; Rouse, 1976; Viereck, 1982). We infer that those sites which burned several decades ago experienced increases in active layer thickness, similar to those seen at sites which burned in 2012, but organic layer re-establishment subsequently led to active layer thinning, a process that may still be underway at site 5 where if permafrost is present, the active layer is about 150 cm thick. Gibson et al. (2018) found extensive permafrost thaw and rapid thermokarst bog development in peatlands following fire in the study region. We did not observe the same rapid degradation, which may be due to dissimilar hydrological and ground ice conditions, as well as differences between peatlands and forested sites. Our results show that not all permafrost near the southern limits of its distribution is sensitive to fire, and it is clearly important to consider the full range of landscape conditions when modelling the trajectory of permafrost in these areas.

3.6 Conclusion

We conclude that significant permafrost degradation has occurred over the past half century along the Mackenzie Highway, especially at sites with thin organic layers overlying coarse-grained substrates, and preferentially at the southern end of the transect where climate warming has resulted in unfavourable conditions for perennially frozen ground. However, permafrost has persisted beneath at least half of the sites along the transect where it was present in 1962, and probably more than 60%, in spite of climate warming and forest fire. Permafrost remains present today under black spruce (*Picea mariana*) canopies at sites with organic layers generally >40 cm thick overlying fine-grained substrates.

We found a much lower degree of permafrost loss than reported in 1994 (Kwong and Gan, 1994) for the same transect, but a similar percent change in sites with permafrost as has been observed for areal change elsewhere in this region (Quinton et al., 2001). Greater frost table depths were observed at sites affected by fires in 2012 but unlike others (Gibson et al., 2018), we did not observe an increase in the overall rate of degradation at sites impacted by forest fire. However, increases to the frequency and magnitude of fires are likely to affect permafrost more in the future as the climate continues to warm.

This study presents evidence for the persistent nature of warm permafrost under a changing climate, even when disturbed. It underlines the importance of considering all landscape types when modelling future permafrost thaw, and more specifically the effects of heterogeneity of drainage conditions, substrate, and organic layer thicknesses on the fate of permafrost in the upcoming decades.

3.7 References

- ACIA. (2005). *Arctic Climate Impact Assessment*. Cambridge University Press. 1020 pp.
- Alberta Government. (2018). *Historical Fire Database*. <http://wildfire.alberta.ca/resources/historical-data/historical-wildfire-database.aspx>
- AMAP. (2017). *Snow, Water, Ice and Permafrost in the Arctic (SWIPA)*. Oslo, Norway: Arctic Monitoring and Assessment Programme.
- Baltzer, J. L., Veness, T., Chasmer, L. E., Sniderhan, A. E., and Quinton, W. L. (2014). Forests on thawing permafrost: fragmentation, edge effects, and net forest loss. *Global Change Biology*, 20(3), 824-834. <https://doi.org/10.1111/gcb.12349>
- Beilman, D. W., and Robinson, S. D. (2003). Recent permafrost melt along a climatic gradient in western Canada: large-scale peat mapping. In M. Phillips, S. M. Springman, L. U. Arenson, A. A. Balkema (Eds.), *Proceedings of the 8th International Conference on Permafrost* (pp. 61-65). Zurich, Switzerland.
- Brown, R. J. E. (1964). *Permafrost investigations on the Mackenzie Highway in Alberta and Mackenzie district*. Ottawa, Canada, National Research Council Technical Paper No. 175.
- Burn, C. R. (1998). The response (1958-1997) of permafrost and near-surface ground temperatures to forest fire, Takhini River Valley, southern Yukon Territory. *Canadian Journal of Earth Science*, 35(2), 184-199. <https://doi.org/10.1139/e97-105>
- Bush, E., and Lemmen D. S., (eds.). (2019). *Canada's Changing Climate Report*. Government of Canada, Ottawa, ON. <https://changingclimate.ca/CCCR2019/>
- Camill, P. (2005). Permafrost thaw accelerates in boreal peatlands during late-20th century climate warming. *Climatic Change*, 68, 135–152. <https://doi.org/10.1007/s10584-005-4785-y>
- Camill, P., and Clark, J. S. (1998). Climate change disequilibrium of boreal permafrost peatlands caused by local processes. *American Naturalist*, 151(3), 207-222. <https://doi.org/10.1086/286112>
- Connon, R., Devoie, É., Hayashi, M., Veness, T., and Quinton, W. (2018). The influence of shallow taliks on permafrost thaw and active layer dynamics in subarctic Canada. *Journal of Geophysical Research: Earth Surface*, 123(2), 281-297. <https://doi.org/10.1002/2017JF004469>
- Connon, R. F., Quinton, W. L., Craig, J. R., and Hayashi, M. (2014). Changing hydrologic connectivity due to permafrost thaw in the lower Liard River valley, NWT, Canada. *Hydrological Processes*, 28(14), 4163-4178. <https://doi.org/10.1002/hyp.10206>
- Doré, G., Niu, F., and Brooks, H. (2016). Adaptation methods for transportation infrastructure built on degrading permafrost. *Permafrost and Periglacial Processes*, 27(4), 352-364. <https://doi.org/10.1002/ppp.1919>

- Dyrness, C. T. (1982). *Control of depth to permafrost and soil temperature by the forest floor in black spruce/feathermoss communities*. USDA Forest Service Pacific Northwest Forest and Range Experiment Station. <https://www.fs.usda.gov/treearch/pubs/7656>
- Ecological Stratification Working Group. (1996). *A National Ecological Framework for Canada*. Centre for Land and Biological Resources Research, Research Branch, Agriculture and Agri-Food Canada.
- Englefield, P. G. C. (1995). *Rate of permafrost degradation in peatlands* [M.Sc. thesis]. University of Alberta. <https://doi.org/10.7939/R3833N81F>
- Environment Canada. (2018). *Historical climate data*. http://climate.weather.gc.ca/historical_data/search_historic_data_e.html
- Farouki, O. T. (1981). *Thermal properties of soils*. Cold Regions Research and Engineering Lab.
- Fisher, J. P., Estop-Aragónés, C., Thierry, A., Charman, D. J., Wolfe, S. A., Hartley, I. P., Murton, J. B., Williams, M., and Phoenix, G. K. (2016). The influence of vegetation and soil characteristics on active-layer thickness of permafrost soils in boreal forest. *Global Change Biology*, 22, 3127–3140. <https://doi.org/10.1111/gcb.13248>
- Fortier, R., LeBlanc, A. M., Allard, M., Buteau, S., and Calmels, F. (2008). Internal structure and conditions of permafrost mounds at Umiujaq in Nunavik, Canada, inferred from field investigation and electrical resistivity tomography. *Canadian Journal of Earth Science*, 45(3), 367–387. <https://doi.org/10.1139/E08-004>
- Fraser, R. H., Hall, R. J., Landry, R., Lynham, T. J., Raymond, D. A., Lee, B. S., and Li, Z. (2004). Validation and calibration of Canada-wide coarse-resolution satellite burned-area maps. *Photogrammetric Engineering & Remote Sensing*, 70, 451–460. <https://doi.org/10.14358/PERS.70.4.451>
- French, H. M. (2018). *The Periglacial Environment* (4th ed.). John Wiley & Sons, Ltd.
- Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., and Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nature Communications*, 9, 3041. <https://doi.org/10.1038/s41467-018-05457-1>
- GNWT. (2018). *GNWT Dataset Download Categories – Forest Management – Fire History*. Northwest Territories Centre for Geomatics. <http://www.geomatics.gov.nt.ca/dldsoptions.aspx>
- Halsey, L. A., Vitt, D. H., and Zoltai, S. C. (1995). Disequilibrium response of permafrost in Boreal Continental Western Canada to climate change. *Climatic Change*, 30, 57–73. <https://doi.org/10.1007/BF01093225>
- Heginbottom, J. A., Dubreuil, M. A., and Harker, P. A. (1995). Canada – Permafrost. *National Atlas of Canada* (5th ed.). Natural Resources Canada.

- Hilbich, C., Marescot, L., Hauck, C., Loke, M. H., and Mäusbacher, R. (2009). Applicability of electrical resistivity tomography monitoring to coarse blocky and ice-rich permafrost landforms. *Permafrost and Periglacial Processes*, 20(3), 269–284. <https://doi.org/10.1002/ppp.652>
- Jafarov, E. E., Romanovsky, V. E., Genet, H., McGuire, A. D., and Marchenko, S. S. (2013). The effects of fire on the thermal stability of permafrost in lowland and upland black spruce forests of interior Alaska in a changing climate. *Environmental Research Letters*, 8(3), 035030. <http://dx.doi.org/10.1088/1748-9326/8/3/035030>
- James, M., Lewkowicz, A. G., Smith, S. L., and Miceli C. M. (2013). Multi-decadal degradation and persistence of permafrost in the Alaska Highway corridor, northwest Canada. *Environmental Research Letters* 8(4): 045013. <http://dx.doi.org/10.1088/1748-9326/8/4/045013>.
- Johnston, G. H., Ladanyi, B., Morgenstern, N. R., and Penner, E. (1981). Engineering Characteristics of Frozen and Thawing Soils. In G. H. Johnston (ed.), *Permafrost Engineering Design and Construction*. John Wiley & Sons.
- Jorgenson, M. T., Racine, C. H., Walters, J. C., and Osterkamp, T. E. (2001). Permafrost degradation and ecological changes associated with a warming climate in central Alaska. *Climatic Change*, 48(4), 551-579. <https://doi.org/10.1023/A:1005667424292>
- Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., O'Donnell, J., Schuur, E. A. G., Kanevskiy, M., and Marchenko, S. (2010). Resilience and vulnerability of permafrost to climate change. *Canadian Journal of Forest Research*, 40(7), 1219-1236. <https://doi.org/10.1139/X10-060>
- Kasischke, E. S., and Stocks, B. J. (2000). *Fire, Climate Change, and Carbon Cycling in the Boreal Forest*. Springer-Verlag.
- Kasischke, E. S., Verbyla, D. L., Rupp, T. S., McGuire, A. D., Murphy, K. A., Jandt, R., Barnes, J. L., Hoy, E. E., Duffy, P. A., Calef, M., and Turetsky, M. R. (2010). Alaska's changing fire regime - implications for the vulnerability of its boreal forests. *Canadian Journal of Forest Research*, 40(7), 1313-1324. <https://doi.org/10.1139/X10-098>
- Kokelj, S. V., and Burn, C. R. (2003). Ground ice and soluble cations in near-surface permafrost, Inuvik, Northwest Territories, Canada. *Permafrost and Periglacial Processes*, 14(3), 275-289. <https://doi.org/10.1002/ppp.458>
- Kwong, Y. J., and Gan, T. Y. (1994). Northward migration of permafrost along the Mackenzie Highway and climatic warming. *Climatic Change*, 26(4), 399-419. <https://doi.org/10.1007/BF01094404>
- Lantz, T. C., and Kokelj, S. V. (2008). Increasing rates of retrogressive thaw slump activity in the Mackenzie Delta region, NWT, Canada. *Geophysical Research Letters*, 35(6). <https://doi.org/10.1029/2007GL032433>

- Lewkowicz, A. G., Etzelmüller, B., and Smith, S. L. (2011). Characteristics of discontinuous permafrost based on ground temperature measurements and electrical resistivity tomography, southern Yukon, Canada. *Permafrost and Periglacial Processes*, 22(4), 320-342. <https://doi.org/10.1002/ppp.703>
- Loke, M. H. (2001). *Electrical imaging surveys for environmental and engineering studies, a practical guide to 2-D and 3-D surveys*. RES2DINV Manual, IRIS Instruments.
- Loke, M. H., Acworth, I., and Dahlin, T. (2003). A comparison of smooth and blocky inversion methods in 2D electrical imaging surveys. *Exploration Geophysics*, 34(3), 182–187. <https://doi.org/10.1071/EG03182>
- Mackay, J. R. (1995). Active layer changes (1968 to 1993) following the forest-tundra fire near Inuvik, N.W.T., Canada. *Arctic and Alpine Research*, 27(4), 323-336.
- McGuire, A. D., Hayes, D. J., Kicklighter, D. W., Manizza, M., Zhuang, Q., Chen, M., Follows, M. J., Gurney, K. R., McClelland, J. W., Melillo, J. M., and Peterson, B. J. (2010). An analysis of the carbon balance of the Arctic Basin from 1997 to 2006. *Tellus B: Chemical and Physical Meteorology*, 62(5), 455-474. <https://doi.org/10.1111/j.1600-0889.2010.00497.x>
- Nelson, F. E., Anisimov, O. A., and Shiklomanov, N. I. (2002). Climate change and hazard zonation in the Circum-Arctic permafrost regions. *Natural Hazards*, 26, 203–225. <https://doi.org/10.1023/A:1015612918401>
- Nossov, D. R., Jorgenson, M. T., Kielland, K., and Kanevskiy, M. Z. (2013). Edaphic and microclimatic controls over permafrost response to fire in interior Alaska. *Environmental Research Letters*, 8(3), 035013. <https://doi.org/10.1088/1748-9326/8/3/035013>
- O'Donnell, J. A., Harden, J. W., McGuire, A. D., Kanevskiys, M. Z., Jorgenson, M. T., and Xu, X. (2011). The effect of fire and permafrost interactions on soil carbon accumulation in an upland black spruce ecosystem of interior Alaska: implications for post-thaw carbon loss. *Global Change Biology*, 17(3), 1461-1474. <https://doi.org/10.1111/j.1365-2486.2010.02358.x>
- Quinton, W. L., Hayashi, M., and Chasmer, L. E. (2009). Peatland hydrology of discontinuous permafrost in the Northwest Territories: overview and synthesis. *Canadian Water Resources Journal*, 34(4), 311-328. <https://doi.org/10.4296/cwrj3404311>
- Quinton, W. L., Hayashi, M., and Chasmer, L. E. (2011). Permafrost-thaw-induced landcover change in the Canadian subarctic: Implications for water resources. *Hydrological Processes*, 25(1), 152–158. <https://doi.org/10.1002/hyp.7894>
- Robinson, S. D., and Moore, T. R. (2000). The influence of permafrost and fire upon carbon accumulation in high boreal peatlands, northwest territories, Canada. *Arctic, Antarctic, and Alpine Research*, 32(2), 155–166. <https://doi.org/10.1080/15230430.2000.12003351>

- Romanovsky, V. E., Smith, S. L., and Christiansen, H. H. (2010). Permafrost thermal state in the polar Northern Hemisphere during the international polar year 2007–2009: a synthesis. *Permafrost and Periglacial Processes*, 21(2), 106–116. <https://doi.org/10.1002/ppp.689>
- Rouse, W. R. (1976). Microclimatic changes accompanying burning in subarctic lichen woodland. *Arctic and Alpine Research*, 8(4), 357–376.
- Schuur, E. A. G., Bockheim, J., Canadell, J. G., Euskirchen, E., Field, C. B., Goryachkin, S. V., Hagemann, S., Kuhry, P., Lafleur, P. M., Lee, H., Mazhitova, G., Nelson, F. E., Rinke, A., Romanovsky, V. E., Shiklomanov, N., Tarnocai, C., Venevsky, S., Vogel, J. G., and Zimov, S. A. (2008). Vulnerability of permafrost carbon to climate change: implications for the global carbon cycle. *BioScience*, 58(8), 701–714. <https://doi.org/10.1641/B580807>
- Shur, Y. L., and Jorgenson, M. T. (2007). Patterns of permafrost formation and degradation in relation to climate and ecosystems. *Permafrost and Periglacial Processes*, 18(1), 7–19. <https://doi.org/10.1002/ppp.582>
- Smith, S. L., and Burgess, M. (2004). *Sensitivity of permafrost to climate warming in Canada*. Bulletin of the Geological Survey of Canada (No. 579), Natural Resources Canada.
- Smith, S. L., Burgess, M. M., Riseborough, D., and Nixon, F. M. (2005). Recent trends from Canadian permafrost thermal monitoring network sites. *Permafrost and Periglacial Processes*, 16(1), 19–30. <https://doi.org/10.1002/ppp.511>
- Smith, S. L., Riseborough, D. W., and Bonnaventure, P. P. (2015). Eighteen-year record of forest fire effects on ground thermal regimes and permafrost in the central Mackenzie Valley, NWT, Canada. *Permafrost and Periglacial Processes*, 26(4), 289–303. <https://doi.org/10.1002/ppp.1849>
- Smith, S.L., Romanovsky, V.E., Lewkowicz, A.G., Burn, C.R., Allard, M., Clow, G.D., Yoshikawa, K., and Throop, J. (2010). Thermal state of permafrost in North America—a contribution to the International Polar Year. *Permafrost and Periglacial Processes*, 21(2), 117–135. <http://dx.doi.org/10.1002/ppp.690>
- Turetsky, M. R., Wieder, R. K., Halsey, L., and Vitt, D. H. (2002). Current disturbance and the diminishing peatland carbon sink. *Geophysical Research Letters*, 29(11), 1–4. <https://doi.org/10.1029/2001GL014000>
- Viereck, L. A. (1982). Effects of fire and firelines on active layer thickness and soil temperatures in interior Alaska. In *Proceedings of the 4th Canadian Permafrost Conference, The Roger J. E. Brown Memorial Volume* (pp. 123–134). National Research Council of Canada.
- Vincent, L. A., Zhang, X., Brown, R. D., Feng, Y., Mekis, E., Milewska, E. J., Wan, H., Wang, X. L. (2015). Observed trends in Canada’s climate and influence of low-frequency variability modes. *Journal of Climate*, 28(6), 4545–4560. <https://doi.org/10.1175/JCLI-D-14-00697.1>

- Way, R. G., Lewkowicz, A. G., and Zhang, Y. (2018). Characteristics and fate of isolated permafrost patches in coastal Labrador, Canada. *The Cryosphere*, 12, 2667-2688. <https://doi.org/10.5194/tc-12-2667-2018>
- Williams, P. J., and Smith, M. W. (Eds.). (1989). The Ground Thermal Regime. In *The Frozen Earth: Fundamentals of Geocryology* (pp. 83-121). Cambridge University Press.
- Wolfe, S. A., Duchesne, C., Kokelj, S. V., Stevens, C. W., Gaanderse, A., Houben, A. J., and D'Onofrio, R. E. (2011). *Report on 2010–2011 permafrost investigations in the Yellowknife Area, Northwest Territories*. Geological Survey of Canada.
- Wolfe, S. A., Morse, P. D., Hoeve, T. E., Sladen, W. E., Kokelj, S. V., and Arenson, L. U. (2015). Disequilibrium permafrost conditions on NWT Highway 3. In *Proceedings of the 68th Canadian Geotechnical Conference & 7th Canadian Permafrost Conference*. Canadian Geotechnical Society.
- Woo, M. K. (1990). Permafrost hydrology. In T. D. Prowse, C. Simon, L. Ommanney (Eds.), *Northern Hydrology, Canadian Perspectives, Proceedings of the Northern Hydrology Symposium* (pp. 63-76). National Hydrology Research Institute.
- Wotton, B.M., Nock, C.A., and Flannigan, M.D. (2010). Forest fire occurrence and climate change in Canada. *International Journal of Wildland Fire*, 19(3), 253-271. <https://doi.org/10.1071/WF09002>
- Yoshikawa, K., Bolton, W. R., Romanovsky, V. E., Fukuda, M., and Hinzman, L. D. (2002). Impacts of wildfire on the permafrost in the boreal forests of Interior Alaska. *Journal of Geophysical Research: Atmospheres*, 107(D1), FFR 4-1-FFR 4-14. <https://doi.org/10.1029/2001JD000438>
- Zhang, Y., Chen, W. J., and Riseborough, D. W. (2008). Transient projections of permafrost distribution in Canada during the 21st century under scenarios of climate change. *Global and Planetary Change*, 60(3-4), 443-456. <https://doi.org/10.1016/j.gloplacha.2007.05.003>
- Zhang, Y., Wolfe, S. A., Morse, P. D., Olthof, I., and Fraser, R. H. (2015). Spatiotemporal impacts of wildfire and climate warming on permafrost across a subarctic region, Canada. *Journal of Geophysical Research: Earth Surface*, 120(11), 2338-2356. <https://doi.org/10.1002/2015JF003679>

CHAPTER 4: FIELD AND LABORATORY INVESTIGATION OF ELECTRICAL RESISTIVITY-TEMPERATURE RELATIONSHIPS, SOUTHERN NORTHWEST TERRITORIES

Abstract: DC electrical resistivity tomography (ERT) is increasingly being used to delimit permafrost bodies, but a threshold value to differentiate between frozen and unfrozen ground is needed for data interpretation. Here we compare resistivities measured in peat and silts across a range of temperatures in both the field and the laboratory. ERT surveys were conducted at four sites with instrumented boreholes in the southern NWT (60.9-62.5°N). In the laboratory, remoulded soil samples were wetted to field moisture content, frozen in closed containers, and resistivity was measured across a range of temperatures during thaw. Between 0°C and -0.5°C, the resistivity of peat ranged from approximately 3000-6000 Ωm . For frozen silt, maximum resistivities were about 1000 Ωm while the lowest was 85 Ωm , well below the value typically used to delimit permafrost. The relationship between resistivity and temperature allowed a two-dimensional temperature field to be generated from the tomograms, but predictions were not always consistent due to intra-site differences in moisture content and the presence or absence of segregated ice. This study illustrates the great range of resistivity values possible at temperatures close to 0°C, and demonstrates the importance of validating ERT results for warm permafrost.

4.1 Introduction

Increases in global air temperatures and precipitation over the next century are expected to be amplified in northern latitudes and to severely impact permafrost (AMAP, 2017; Zhang et al., 2008). Future changes will add to permafrost thaw that has occurred over the past several decades, especially for warm permafrost (e.g. Gibson et al., 2018; James et al., 2013; Smith et al., 2010). Mapping and delimiting permafrost is vital to understand how it will be impacted by climate change, for land and infrastructure managers, and for residents and governments of northern regions.

Geophysical measurements can provide a non-invasive, two- or three-dimensional portrayal of permafrost conditions. In particular, electrical resistivity tomography (ERT) has been widely used to delimit permafrost due to the unique electrical properties of frozen ground (e.g. Douglas et al., 2016; Lewkowicz et al., 2011; Way et al., 2018). Interpretation of ERT

results, however, is partly subjective as the resistivity boundary between unfrozen and frozen ground is variable (Table 4-1; Hauck, 2013). This can be especially challenging in the discontinuous permafrost zone, where frozen soil at temperatures between -1 and 0°C can exhibit resistivities that overlap with those of unfrozen soil (Hoekstra et al., 1974). Furthermore, much of the extant literature focuses on relatively cold, ice-rich permafrost with high resistivities. This could lead to misinterpretation of ERT results for warm or thawing discontinuous permafrost. The objective of this study was to improve the interpretation of ERT results by comparing measurements made in the field and in the laboratory for soils at a range of temperatures close to 0°C.

Table 4-1: Resistivity values used in the literature for the boundary between frozen and unfrozen soils. Note: this table was not included in the original publication.

Study	Location	Soil Type	Resistivity (Ωm)
Lewkowicz et al. 2011	Yukon, Canada	Peat	1000
Lewkowicz et al. 2011	Yukon, Canada	Peat	300
McClymont et al. 2013	NWT, Canada	Peat	1000
Sjoberg et al. 2015	Sweden	Peat	1000 - 1700
Way et al. 2018	Labrador, Canada	Peat	300 - 800
Seguin and Frydecki 1994	Quebec, Canada	Fine-grained sediment	1000
Yoshikawa et al. 2006	Alaska, USA	Silt	8000
Douglas et al. 2008	Alaska, USA	Silt	450 – 800
Fortier et al. 2008	Nunavik, Canada	Glaciomarine sediment	1000
Douglas et al. 2016	Alaska, USA	Silt	1000
Way and Lewkowicz 2015	Labrador, Canada	Medium to fine grained sediment	300 - 400
Way and Lewkowicz 2015	Labrador, Canada	Medium to fine grained sediment	800 - 1000
Lewkowicz et al. 2011	Yukon, Canada	Sand and gravel	700 - 800
Yoshikawa et al. 2006	Alaska, USA	Massive ice	18,000

4.2 Field sites

We set up four monitoring sites within 1 km of the Mackenzie Highway in 2015 and 2016 between Kakisa, NT (60.9°N) and Yellowknife, NT (62.5°N) (Figure 4-1). Climate in the region is subarctic and continental, with mean annual air temperature (MAAT) for 1981-2010 of -2.5°C at Hay River and -4.3°C at Yellowknife (Environment Canada, 2019). Annual precipitation (1981-2010) is 336 mm at Hay River and 289 mm at Yellowknife, of which 41% and 55% falls as snow, respectively (Environment Canada, 2019). Climate in the region has warmed by ~2°C over the last half century (Holloway and Lewkowicz, 2020).

The study transect traverses the taiga plains and the taiga shield ecozones. The former

comprises flat to gently rolling terrain covered by glaciolacustrine deposits and till. The taiga shield is comprised of undulating to hilly exposed bedrock, with thin till and wetlands in low-lying areas. The surface cover is primarily northern boreal forest, but two of our sites (SS33-9a and ZF20-6b-LK) were burned by wildfire in 2014. The southern part of the transect falls into the zone of sporadic discontinuous permafrost (Heginbottom et al., 1995) where thicknesses are less than 30 m (Smith et al., 2015), and mean annual ground temperatures are above -1°C (Smith et al., 2010). Near its southern margin, permafrost occurs mostly in peatlands, and has been degrading over the last century (Beilman et al., 2003; Smith et al., 2010). At the boundary with extensive discontinuous permafrost near Yellowknife, permafrost is less than 50 m thick and generally warmer than -2°C (Wolfe et al., 2015).

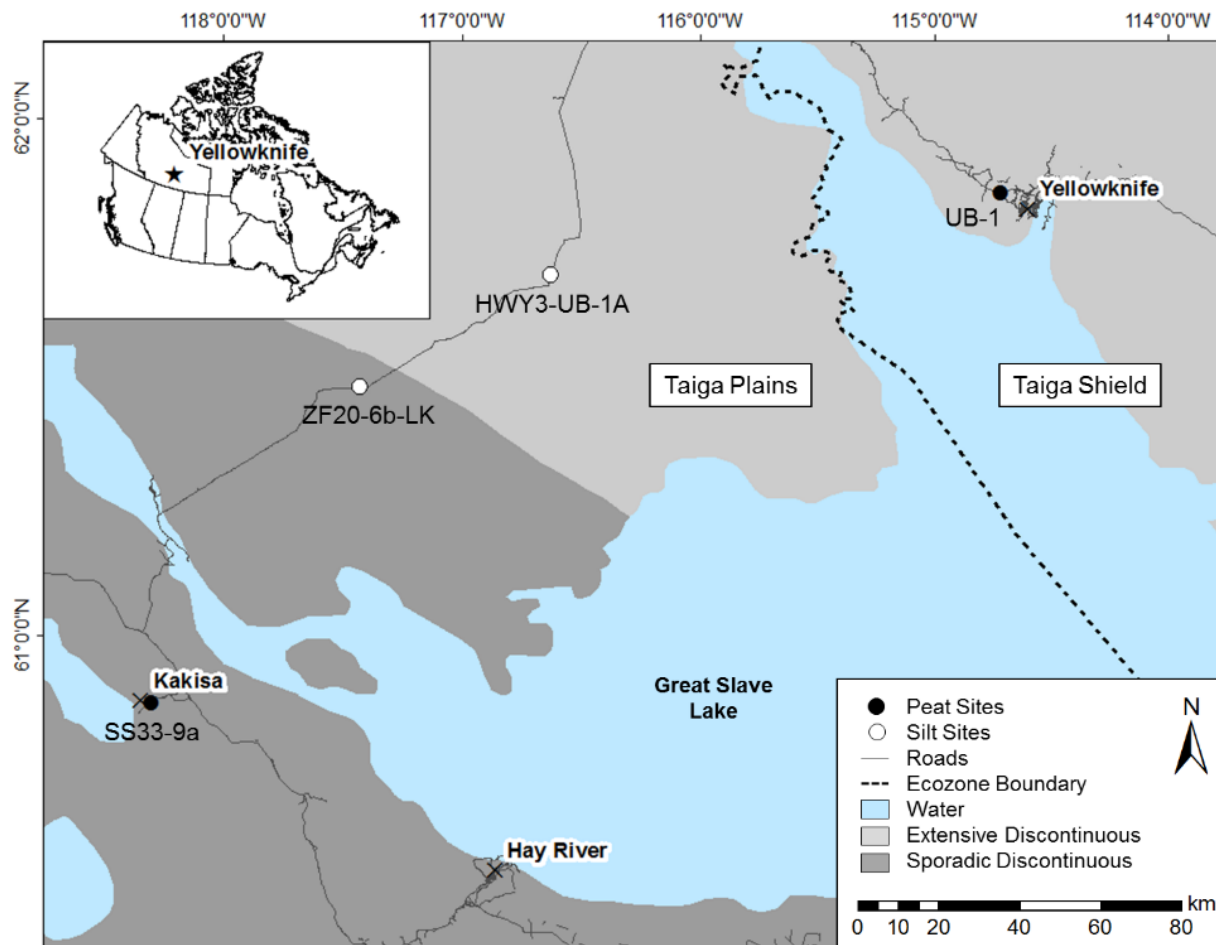


Figure 4-1: Study transect along the Mackenzie Highway showing site locations and soil type, permafrost zonation (from Heginbottom et al., 1995), and ecozones. Inset map shows the location of Yellowknife, NT.

4.3 Methods

4.3.1 Field measurements

Ground temperatures were measured in boreholes drilled using the water-jet technique and cased in 50 mm I.D. ABS pipe to depths between 1.5-7.5 m (Table 4-2). Temperatures were recorded at 2-hourly intervals between May 2015 and September 2018 using Onset Corporation (USA) Hobo Pro U12-008 loggers with four external thermistors (accuracy $\pm 0.25^{\circ}\text{C}$; precision $\pm 0.03^{\circ}\text{C}$) and Hobo Pro U23-003 loggers with two external thermistors (accuracy $\pm 0.21^{\circ}\text{C}$; precision $\pm 0.02^{\circ}\text{C}$) (Table 4-2).

Table 4-2: Field site information.

Site	Temp. ^a , Depth, Date	ERT Survey Dates	Core Depths	GS ^b (μm)	MC ^c (%)
UB-1	-0.5°C, 7.4 m, 13/9/18	7/6/15, 11/8/15, 31/5/16, 15/9/16, 7/9/17, 13/9/18 (160 m)	145 cm	Peat	640
HWY3-UB-1A	-0.5°C, 1.5 m, 10/9/18	18/5/16, 9/9/16, 3/9/17, 10/9/18 (160 m)	153 cm	33	73
ZF20-6b-LK	-0.2°C, 3.2 m, 5/9/16	12/6/15, 18/8/15, 13/5/16, 5/9/16, 2/9/17, 6/9/18 (160 m); 15/5/16, 5/9/16 (40 m)	120 cm	63	43
SS33-9a	-0.14°C, 3.1 m, 3/9/18	17/6/15, 16/8/15, 17/5/16, 1/9/16, 29/8/17, 3/9/18 (160 m); 1/9/16, 29/8/17, 3/9/18 (40 m)	310 cm	Peat	739

^a mean temperature measured at the deepest thermistor on the day of ERT surveys (Figure 4-3); ^b mean grain size; ^c gravimetric moisture content of the cores. Note that the HWY3-UB-1A and ZF20-6b-LK cores contained excess ice.

An ABEM LS multi-electrode, two-dimensional system was used to conduct ERT surveys at all four sites (Table 4-2). Steel electrodes were arranged in a Wenner array, with a minimum spacing of either 2 m and a depth of investigation of 24 m for 160 m long surveys, or a 0.25 m spacing and depth of 6 m for 40 m surveys. Tomograms were produced using the robust inversion method in the software package RES2DINV, which performs smoothness-constrained inversions using finite difference forward modelling (Loke, 2001). Iteration was carried out until the RMS error fell below 5%, or to the fifth iteration, whichever occurred first. Model sensitivity was calculated in RES2DINV (Loke, 2003), and model blocks with sensitivity less than 0.1 were excluded (Way et al., 2018).

4.3.2 Permafrost sampling and laboratory measurements

Soil samples from the active layer and upper permafrost were collected in 2016 using a motorized SIPRE corer with a 75 mm diameter core barrel. Cores were kept frozen and shipped back to Ottawa for laboratory analysis. Samples were thawed, and then dried in an oven at 105°C for 24 hours to calculate the gravimetric moisture content. All samples from a given site were then homogenized and remoulded into a single sample. Samples from each site were rehydrated to field moisture content with deionised water and then refrigerated for three days. Each sample was placed into a closed 15 cm diameter PVC container at field bulk density. Four 12/2-gauge copper wires were inserted as electrodes 5 cm below the soil surface, in a Schlumberger array, with 3 cm spacing. Hobo Pro U23-003 loggers with two external thermistors were inserted to measure temperature. The containers were placed in an environmental test chamber (model RTC-16P made by Burnco Technologies Inc.) with a nominal temperature control of $\pm 1^\circ\text{C}$ and cooled to -20°C . The chamber was then warmed to -10 , -5 , -2.5 , -1.5 , -1 , -0.5 , 0.5 , and 5°C , the samples were allowed to equilibrate at each nominal temperature, and the ABEM system was used to measure resistivity while the Hobo loggers provided actual soil temperature.

4.3.3 Data processing

Resistivity measurements were extracted from inverted ERT field surveys for model blocks which bridged the temperature measurement depths in the borehole and at the location of the borehole along each ERT profile (± 3 m horizontal distance) (Table 4-3). Linear interpolation was used between temperature measurement depths within the boreholes to infer the temperature

at the depth of the centre of the model block. Least-squares log-linear relationships were established between these temperatures and resistivity at three of the four sites. Site HWY3-UB-1A was excluded because there were too few data points (Table 4-3). Values above and below 0°C were treated separately, generating two equations per site. Inverted resistivity values from September 2016 (site ZF20-6b-LK) or September 2018 (sites UB-1 and SS33-9a) were converted into predicted temperatures using the equations in Table 4-3. The minimum resistivity value recorded below 0°C determined which equation was used for the conversion (Table 4-3). The Kriging method in Surfer (Version 16, Golden Software) was used to create two-dimensional contour maps of the predicted temperatures.

Table 4-3: Relationship between resistivity and temperature.

Site	Minimum Resistivity (Ωm)	Temperature	n	Equation ^a	R ²
UB-1	224	<0°C	6	$y = -3.690x + 2.509$	0.93
		>0°C	8	$y = -0.136x + 2.031$	0.59
ZF20-6b-LK	84	<0°C	10	$y = -0.102x + 2.210$	0.96
		>0°C	12	$y = -0.089x + 2.362$	0.57
SS33-9a	354	<0°C	6	$y = -0.932x + 2.933$	0.92
		>0°C	22	$y = -0.246x + 3.164$	0.52

^aIn the equation, y is the logarithm (base 10) of resistivity (Ωm) and x is temperature (°C)

4.4 Results and Discussion

4.4.1 Variation in resistivity with temperature

Figure 4-2 shows the relationship between resistivity and temperature as measured in the field and laboratory. Minimum temperatures in the boreholes during the time of ERT surveys were -0.2 to -0.8°C in May and -0.2 to -0.7°C in August-September. Maximum temperatures during the field surveys reached 12.1°C in the active layer. In the laboratory experiments, the lowest temperature at which resistivities could be measured varied from -19°C for silts to -1°C for high ice-content peat. Below -1°C in peat, resistivities were so high that the instrument regarded them as an absence of electrical contact.

As expected, the rate of change of resistivity vs. temperature increases sharply at or very close to 0°C at all sites (Figure 4-2), giving an order of magnitude difference in resistivity values between 0.5°C and -0.5°C (Figure 4-2). This reflects initial ice nucleation and the impact of progressive decrease in unfrozen moisture content as the temperature falls below 0°C (Hoekstra et al., 1974). Doetsch et al. (2015) found similar results for saturated sand and gravel. Resistivity values climb to between 3000 and 6000 Ωm at -0.5°C for the two peat sites whereas the values at -0.5°C are closer to 100 Ωm for the two silt sites in the laboratory. Field resistivities at silt site ZF20-6b-LK are generally higher, ranging from 100 to 1000 Ωm at temperatures between -0.15 and -0.2°C. The lowest resistivity value at a temperature below 0°C is 85 Ωm at -0.14°C at site ZF20-6b-LK (silt). At temperatures > 0°C, resistivities range between 10 and 100 Ωm , there is little or no slope and measurements from the field are generally higher than those in the laboratory. It is important to note that the geochemical composition of the soil water determines the freezing-point depression (Andersland and Ladanyi, 2004), and impacts the unfrozen moisture content (Fisher et al., 2019). Although we did not measure the soil water chemistry, we assume the solute content to be low since the soils are not marine clays. Consequently, there is likely only a very small freezing-point depression.

4.4.2 ERT tomograms and temperature contour maps

At site UB-1, a continuous high resistivity body extends across the profile, with low resistivities in the near-surface in the last 40 m of the transect (Figure 4-3a). At site SS33-9a, a near-surface high resistivity body overlies a low resistivity layer which extends to the base of the tomogram (Figure 4-3b). Low resistivity anomalies in the near-surface in the middle of the transect correspond to areas of thermokarst. At site ZF20-6b-LK, a high resistivity body is present at depth in the first half of the transect while low values extend across the remainder of the profile (Figure 4-3c). For this interpretation, the minimum resistivity value recorded at a temperature below 0°C in the laboratory or in the field was used as the threshold between frozen and unfrozen soil (Table 4-3). Using this value, there is 17 m, 12 m, and 20 m of permafrost at sites UB-1, SS33-9a, and ZF20-6b-LK respectively.

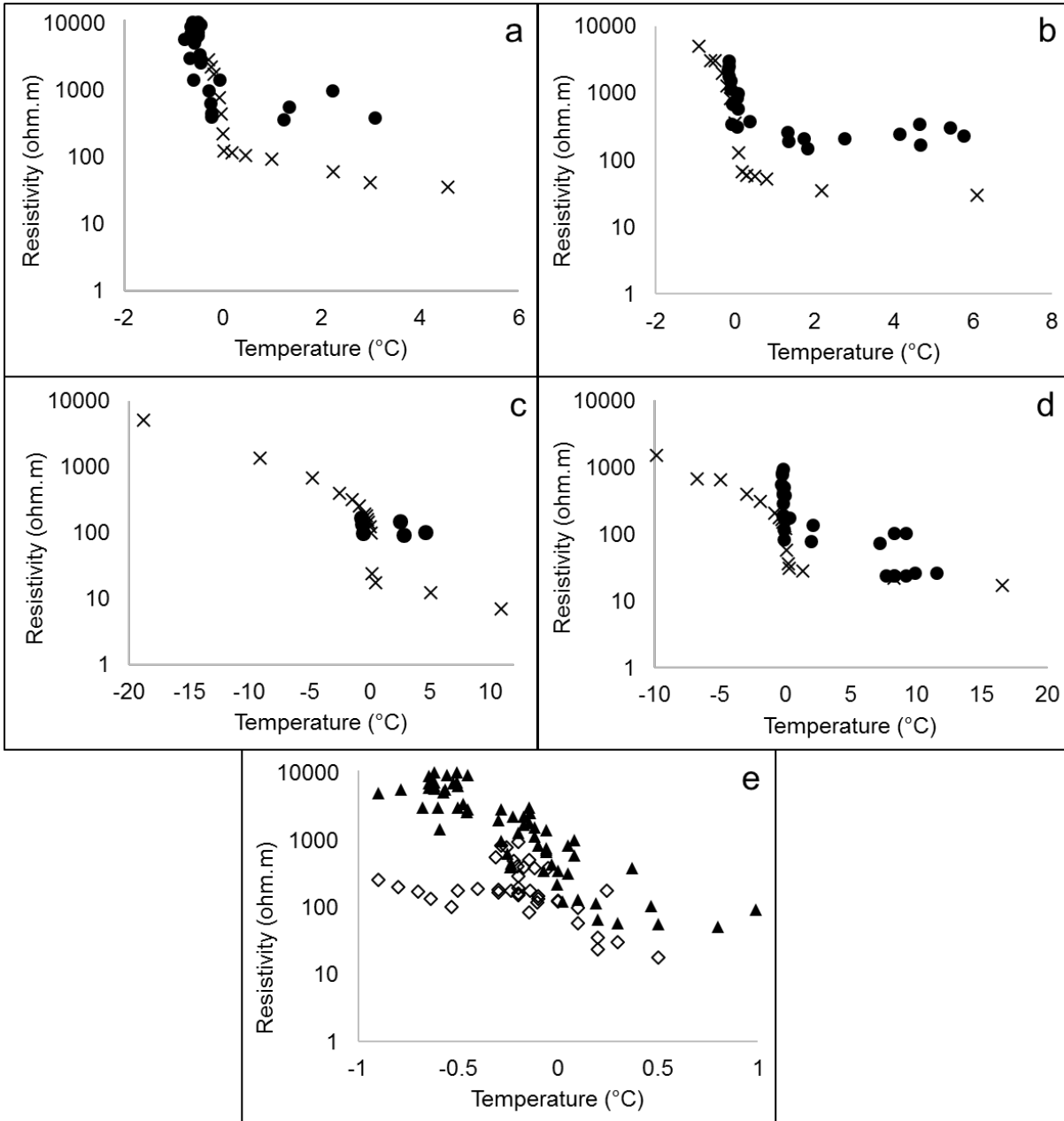


Figure 4-2: Relationship between resistivity and temperature for measurements in the field (dots) and laboratory (crosses) at (a) UB-1 (peat), (b) SS33-9a (peat), (c) HWY3-UB-1A (silt), (d) ZF20-6b-LK (silt), and (e) all sites for temperatures between -1°C and 1°C differentiated as peat (triangles) and silt (diamonds). Note the varying temperature scales.

Predicted temperatures for each site are contoured in Figure 4-4. For site UB-1, predicted temperatures range from -1.2°C to 30.7°C , with the lowest temperatures in a body at 5 m depth between 30 and 60 m along the profile (Figure 4-4a). Temperatures range from -0.2°C to 18.6°C at site SS33-9a (Figure 4-4b), and from -0.4°C to 13.6°C at site ZF20-6b-LK (Figure 4-4c).

4.4.3 Resistivity and temperature measurements in the field and laboratory

The steep increase in resistivity at 0°C corresponds to previous observations (e.g. Hoekstra et al., 1974). However, in some instances we observed this step change in the laboratory at temperatures around 0.2°C (Figure 4-2). This probably reflects the relative positioning of the temperature sensor and electrodes so that the soil thawed at the former before the latter.

We observed a much greater increase in resistivity at 0°C for peat sites compared to silt sites (Figure 4-2e), because the former have higher porosity and moisture content, and therefore ice content, which dramatically increases resistivity (Hoekstra et al., 1974). This shows the importance of validating substrate for interpreting ERT results. Compared to the results from Hoekstra et al. (1974), we observed lower resistivities for our silt sites at temperatures above 0°C and a sharper change at 0°C . We again believe this is due to differences in moisture: we used field moisture content while Hoekstra et al. (1974) saturated their samples. Finally, we observed a difference in resistivity values $<0^{\circ}\text{C}$ between the laboratory and field, with field measurements higher than laboratory measurements at every site. We believe that this is due to differences in soil structure, in particular the presence of segregated ice (observed in the cores), which would increase the resistivity in the field but which we did not replicate in the laboratory experiments.

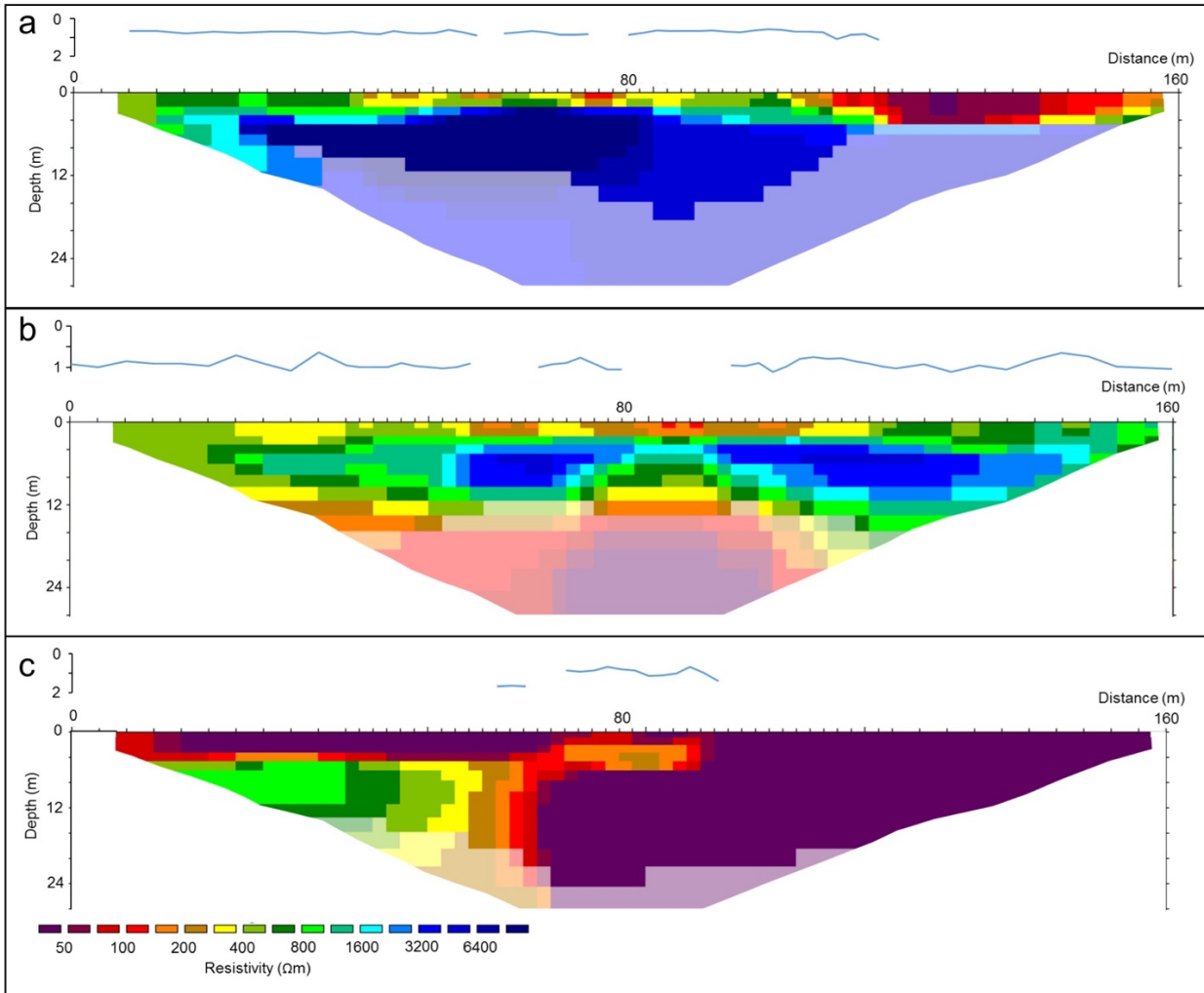


Figure 4-3: 160 m long ERT surveys and frost table depths (blue lines) for (a) site UB-1, (b) site SS33-9a and (c) site ZF20-6b-LK. Surveys (a) and (b) were undertaken in September 2018 and (c) in September 2016. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded. The minimum resistivity value recorded at a temperature below 0°C in the laboratory or in the field was used as the threshold between frozen and unfrozen soil. Note: SS33-9a profile is the 4th iteration, RMS error 4.9%; UB-1 profile is the 5th iteration, RMS error 5.4%; and ZF20-6b-LK profile is the 3rd iteration, RMS error 3.9%. Frost table depths were measured at 2-4 m intervals and blank areas indicate no frost table found within 110 cm of the surface.

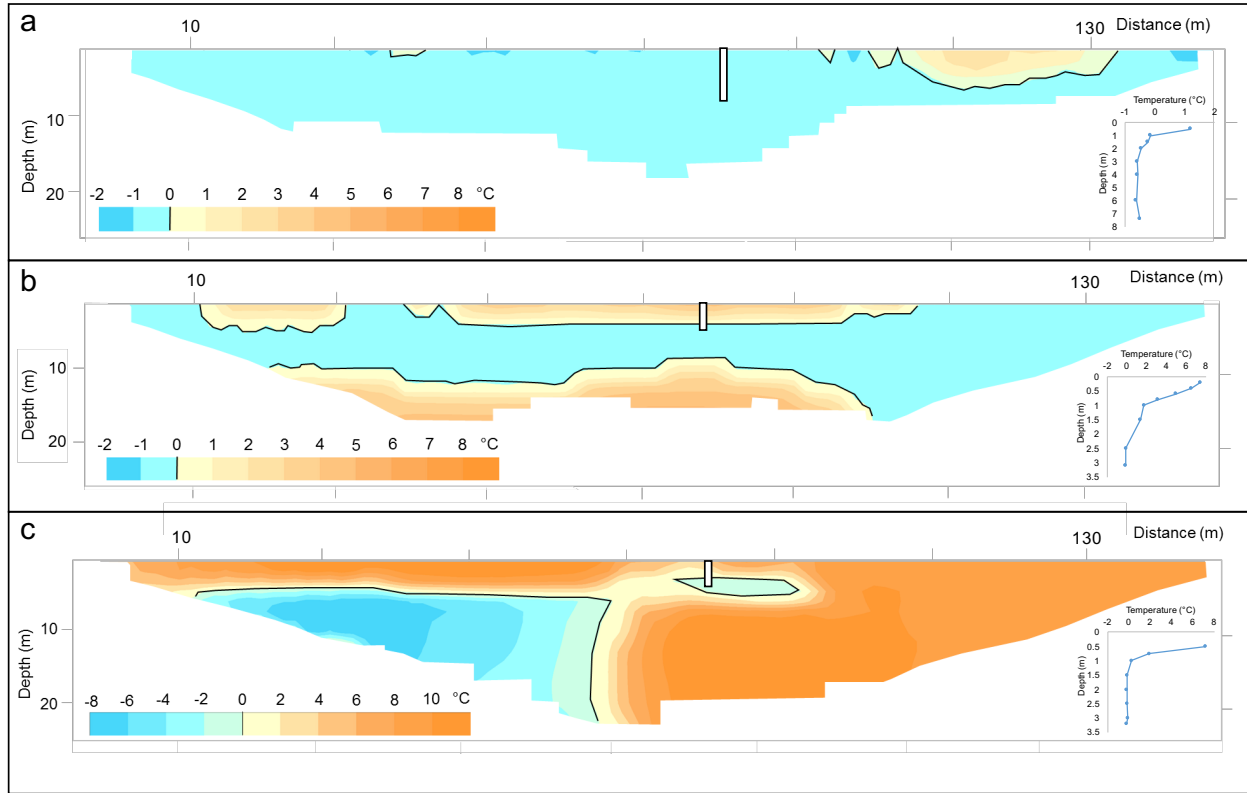


Figure 4-4: Two-dimensional contour plots of predicted temperatures at UB-1 (a), SS33-9a (b), and ZF20-6b-LK (c). Areas where resistivity measurements had sensitivity values below 0.1 have been blanked out. Note: inset temperature profiles are from the boreholes (white box) on the day of the ERT surveys (Table 4-2). They were not included in the original publication.

4.4.4 Comparison of ERT tomograms and temperature contour plots

As far as we know, this is the first attempt to use resistivity-temperature relationships to predict ground temperatures (Figure 4-4). The predicted values correspond well to conditions observed in the field at UB-1 and SS33-9a for temperature (Table 4-2) and frost table depths (Figure 4-3). The relationship was not able to capture intra-site differences in moisture content or the presence of segregated ice, resulting in predicted temperatures at ZF20-6b-LK being lower than those observed in the field (Figure 4-4c, Table 4-2). However, the anomaly in the midsection of ZF20-6b-LK is correctly predicted as frozen ground, despite its low resistivity values (Figure 4-4c). The temperatures in the borehole at 80 m were between 0°C and -0.2°C at depths of 1.5 to 3.2 m. This emphasizes the need to consider the potential for lower resistivity thresholds for permafrost close to 0°C.

4.5 Conclusions

We have shown that frozen ground can exhibit a wide range of resistivities close to 0°C. Between 0°C and -0.5°C, the resistivity of peat ranges from approximately 3000-6000 Ωm . For silt, the range is from 1000 Ωm to 85 Ωm , the latter being much lower than other published observations (Table 4-1). We conclude that low resistivity thresholds need to be considered for warm or thawing permafrost.

Resistivity-temperature relationships differ for peat and silt substrates, underlining the importance of stratigraphic knowledge when using ERT to evaluate permafrost conditions. Using the relationship between resistivity and temperature we were able to delimit permafrost bodies in two-dimensions, resulting in a more thorough understanding of the site conditions. However, the predictions were affected by intra-site differences in moisture content and segregated ice.

This study presents further evidence that ERT surveys can be used to assess permafrost boundaries and conditions, but emphasizes the importance of validating the results, especially for warm permafrost.

4.6 References

- AMAP. (2017). *Snow, Water, Ice and Permafrost in the Arctic (SWIPA)*. Oslo, Norway: Arctic Monitoring and Assessment Programme.
- Andersland, O., and Ladanyi, B. (2004). *Frozen Ground Engineering*. 2nd ed. Chichester, UK: John Wiley & Sons.
- Beilman, D. W., and Robinson, S. D. (2003). Recent permafrost melt along a climatic gradient in western Canada: large-scale peat mapping. In M. Phillips, S. M. Springman, L. U. Arenson, A. A. Balkema (Eds.), *Proceedings of the 8th International Conference on Permafrost* (pp. 61-65). Zurich, Switzerland.
- Doetsch, J., Ingeman-Nielsen, T., Christiansen, A. V., Fiandaca, G., Auken, E., and Elberling, B. (2015). Direct current (DC) resistivity and induced polarization (IP) monitoring of active layer dynamics at high temporal resolution. *Cold Regions Science and Technology*, 119, 16-28. <https://doi.org/10.1016/j.coldregions.2015.07.002>
- Douglas, T. A., Jorgenson, M. T., Brown, D. R., Campbell, S. W., Hiemstra, C. A., Saari, S. P., Bjella, K., and Liljedahl, A. K. (2016). Degrading permafrost mapped with electrical resistivity tomography, airborne imagery and LiDAR, and seasonal thaw measurements. *Geophysics*, 81(1), 1JF-Z7. <https://doi.org/10.1190/geo2015-0149.1>
- Douglas, T. A., Jorgenson, M. T., Kanevskiy, M. Z., Romanovsky, V. E., Shur, Y., and Yoshikawa, K. (2008). Permafrost dynamics at the Fairbanks Permafrost Experimental Station near Fairbanks, Alaska. In D. Kane, and K. Hinkle (Eds.), *Proceedings of the Ninth International Conference on Permafrost* (pp. 373-378). University of Alaska Fairbanks.
- Environment Canada. (2019). *Historical climate data*. http://climate.weather.gc.ca/historical_data/search_historic_data_e.html.
- Fisher, D. A., Lacelle, D., and Pollard, W. (2019). A model of unfrozen water content and its transport in icy permafrost soils: Effects on ground ice content and permafrost stability. *Permafrost and Periglacial Processes*, 31(1), 184-199. <https://doi.org/10.1002/ppp.2031>
- Fortier, R., LeBlanc, A. M., Allard, M., Buteau, S., and Calmels, F. (2008). Internal structure and conditions of permafrost mounds at Umiujaq in Nunavik, Canada, inferred from field investigation and electrical resistivity tomography. *Canadian Journal of Earth Science*, 45(3), 367–387. <https://doi.org/10.1139/E08-004>
- Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., and Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nature Communications*, 9, 3041. <https://doi.org/10.1038/s41467-018-05457-1>

- Hauck, C. (2013). New concepts in geophysical surveying and data interpretation for permafrost terrain. *Permafrost and Periglacial Processes*, 24(2), 131-137. <https://doi.org/10.1002/ppp.1774>
- Heginbottom, J. A., Dubreuil, M. A., and Harker, P. A. (1995). Canada – Permafrost. *National Atlas of Canada* (5th ed.). Natural Resources Canada.
- Hoekstra, P., Sellmann, P.V., and Delaney, A.J. (1974). *Airborne resistivity mapping of permafrost near Fairbanks, Alaska*. Cold Regions Research and Engineering Laboratory, Report 324.
- Holloway, J. E., and Lewkowicz, A. G. (2020). Persistence of permafrost following 55 years of climate warming and forest fire in the discontinuous zone, Canada. *Permafrost and Periglacial Processes*, 31(1), 85-96. <https://doi.org/10.1002/ppp.2017>
- James, M., Lewkowicz, A. G. Smith, S. L., and Miceli C. M. (2013). Multi-decadal degradation and persistence of permafrost in the Alaska Highway corridor, northwest Canada. *Environmental Research Letters* 8(4): 045013. <http://dx.doi.org/10.1088/1748-9326/8/4/045013>.
- Lewkowicz, A. G., Etzelmüller, B., and Smith, S. L. (2011). Characteristics of discontinuous permafrost based on ground temperature measurements and electrical resistivity tomography, southern Yukon, Canada. *Permafrost and Periglacial Processes*, 22(4), 320-342. <https://doi.org/10.1002/ppp.703>
- Loke, M.H. (2001). *Electrical imaging surveys for environmental and engineering studies, a practical guide to 2-D and 3-D surveys*. RES2DINV Manual, IRIS Instruments.
- Loke, M. H., Acworth, I., and Dahlin, T. (2003). A comparison of smooth and blocky inversion methods in 2D electrical imaging surveys. *Exploration Geophysics*, 34(3), 182–187. <https://doi.org/10.1071/EG03182>
- McClymont, A. F., Hayashi, M., Bentley, L. R., and Christensen, B. S. (2013). Geophysical imaging and thermal modeling of subsurface morphology and thaw evolution of discontinuous permafrost. *Journal of Geophysical Research: Earth Surface*, 118(3), 1826–1837. <https://doi.org/10.1002/jgrf.20114>
- Seguin, M-K. and Frydecki, J. (1994). Semi-quantitative geophysical investigation of permafrost in northern Québec. *Journal of Applied Geophysics*, 32:73-84. [https://doi.org/10.1016/0926-9851\(94\)90010-8](https://doi.org/10.1016/0926-9851(94)90010-8)
- Sjöberg, Y., Marklund, P., Pettersson, R., and Lyon, S.W. (2015). Geophysical mapping of peatland permafrost. *The Cryosphere*, 9:465-478. <https://doi.org/10.5194/tc-9-465-2015>
- Smith, S. L., Riseborough, D. W., and Bonnaventure, P. P. (2015). Eighteen-year record of forest fire effects on ground thermal regimes and permafrost in the central Mackenzie Valley,

- NWT, Canada. *Permafrost and Periglacial Processes*, 26(4), 289-303. <https://doi.org/10.1002/ppp.1849>
- Smith, S.L., Romanovsky, V.E., Lewkowicz, A.G., Burn, C.R., Allard, M., Clow, G.D., Yoshikawa, K., and Throop, J. (2010). Thermal state of permafrost in North America—a contribution to the International Polar Year. *Permafrost and Periglacial Processes*, 21(2), 117–135. <http://dx.doi.org/10.1002/ppp.690>
- Way, R. G. and Lewkowicz, A. G. (2015). Investigations of discontinuous permafrost in coastal Labrador with DC electrical resistivity tomography. In *Proceedings of the 68th Canadian Geotechnical Conference & 7th Canadian Permafrost Conference*. Canadian Geotechnical Society.
- Way, R. G., Lewkowicz, A. G., and Zhang, Y. (2018). Characteristics and fate of isolated permafrost patches in coastal Labrador, Canada. *The Cryosphere*, 12, 2667-2688. <https://doi.org/10.5194/tc-12-2667-2018>
- Wolfe, S. A., Morse, P. D., Hoeve, T. E., Sladen, W. E., Kokelj, S. V., and Arenson, L. U. (2015). Disequilibrium permafrost conditions on NWT Highway 3. In *Proceedings of the 68th Canadian Geotechnical Conference & 7th Canadian Permafrost Conference*. Canadian Geotechnical Society.
- Yoshikawa, K., Leuschen, C., Ikeda, A., Harada, K., Gogineni, P., Hoekstra, P., Hinzman, L., Sawada, Y., and Matsuoka, N. (2006). Comparison of geophysical investigations for detection of massive ground ice (pingo ice). *Journal of Geophysical Research: Planets*, 111(E06). <https://doi.org/10.1029/2005JE002573>
- Zhang, Y., Chen, W. J., and Riseborough, D. W. (2008). Transient projections of permafrost distribution in Canada during the 21st century under scenarios of climate change. *Global and Planetary Change*, 60(3-4), 443-456. <https://doi.org/10.1016/j.gloplacha.2007.05.003>

Note: the following references were not cited in the original published paper: Andersland and Ladanyi, 2004; Douglas et al., 2008; Fisher et al., 2019; Fortier et al., 2008; McClymont et al., 2013; Seguin and Frydecki, 1994; Sjöberg et al., 2015; Way et al., 2015; Yoshikawa et al., 2006.

CHAPTER 5: CHANGES TO PERMAFROST IN THE 5 YEARS AFTER FOREST FIRE, SOUTHERN NORTHWEST TERRITORIES, CANADA

5.1 Abstract

Increases in the frequency, severity and extent of wildfire, in combination with climate warming, are expected to impact permafrost landscapes, especially in the boreal forest. This study examines permafrost trajectories at 11 monitoring sites located between 60.9-63.1°N in the southern Northwest Territories (Canada) following the extensive fire year of 2014 and more localized fires in 2015. Monitoring from 2015 to 2019 included measurements of ground and air temperature, snow depths, frost table depths, ground subsidence, and repeated direct current electrical resistivity tomography (ERT) surveys. Permafrost change occurred in the near surface (<5 m depth) at all sites. Increases in frost table depths and ground temperatures, and reductions in resistivity, were greater at the burned sites, especially those with low gravimetric moisture contents, coarse soil textures, and thin organic layers (<40 cm). Ground subsidence and thermokarst development were limited by generally low ice contents, but where these developed, water inundation and permafrost thaw occurred, particularly in areas which were severely burned. Snow-depth days, maximum snow depths, and the nival offset were greater at burned sites than unburned ones. Regionally, colder permafrost sites at the northern end of the transect, as well as permafrost under thick organic layers at the southern end, appear resilient to fire (at least in the short term), whereas sites with thin organic layers and coarse textured soils are the most vulnerable to thaw. We conclude that permafrost can persist following forest fire in certain settings, even in the sporadic discontinuous zone, which underlines the importance of considering sites across a range of landscape types, particularly in relation to ground-ice conditions.

5.2 Introduction

Forest fire is a natural, relatively infrequent, phenomenon (Hanes et al., 2019; Kasischke et al., 2010; Wang et al., 2015) affecting permafrost in North America (Gibson et al., 2018; Nossov et al., 2013), China (Li et al., 2019), and Russia (Knorre et al., 2019). Fire has been shown to impact permafrost landscapes by altering the surface energy balance, ground thermal and hydrological regimes, and soil and aquatic biogeochemistry (Gibson et al., 2018; Holloway

et al., 2020; O'Donnell et al., 2011b). Additionally, ice-rich sites may develop thermokarst following fire (Brown et al., 2015; Douglas et al., 2016; Minsley et al., 2016; Nossov et al., 2013). Despite these impacts, permafrost has historically been relatively resilient to fire over the long term, with variables such as active layer thickness, ground temperatures, and soil moisture returning to pre-fire levels after several decades (Burn, 1998; Gibson et al., 2018; Smith et al., 2015; Viereck et al., 2009). Even post-fire thermokarst is followed by autogenic succession of a peat layer and aggradation of permafrost (Jones et al., 2012). However, the growing frequency, severity, and extent of fires in the northern hemisphere, in combination with climate warming, is altering post-fire permafrost response, and will likely continue to do so in the future (de Groot et al., 2013; Hanes et al., 2019; Jafarov et al., 2013; Kasischke et al., 2010; Wotton et al., 2017; Zhang et al., 2015).

Notwithstanding these generalizations about climate and fire-induced permafrost change, more empirical evidence is needed to validate modelling results (Holloway et al., 2020). Most empirical studies of post-fire impacts on lowland permafrost describe rapid development of thermokarst at high ice-content sites (Baltzer et al., 2014; Gibson et al., 2018), while modelling is typically focused on slower change at low ice-content sites (Jafarov et al., 2013; Zhang et al., 2015). Field studies are needed to evaluate the impact of fire on low ice-content sites, which are common in permafrost regions (Olefeldt et al., 2016; O'Neill et al., 2019; 2020). Their thaw could alter hydrological processes and biogeochemical cycling, including carbon fluxes.

Over the last decade, fewer than 15 studies have addressed the impacts of wildfire on permafrost within Canada (Holloway et al., 2020). The temporal behavior of several variables that affect post-fire permafrost dynamics, such as soil moisture, snow depth, burn severity remain poorly known. Furthermore, the majority of post-fire permafrost studies address edaphic characteristics and near-surface impacts, neglecting permafrost conditions at depth. Geophysical techniques such as electrical resistivity tomography (ERT) are now widely used to evaluate permafrost at depth (e.g., Fortier et al., 2008; Way et al., 2018). To our knowledge, however, these techniques have been used only three times post-fire (Douglas et al., 2016; Lewkowicz et al., 2011; Northern Climate Exchange, 2013), and never repeatedly to monitor permafrost change over time.

The goal of this study is to address several key knowledge gaps by examining the impacts of the abnormally intense fire year of 2014, and more localized fires of 2015, in the Northwest

Territories (NWT), Canada. The extensive fires meant that it was possible to access burned sites with a wide range of environmental conditions (Canadian Interagency Forest Fire Center, 2014; Walker et al., 2018b).

5.3 Study Area

Field investigations were carried out from May 2015 to September 2019 along a transect of sites that followed the Mackenzie Highway between Kakisa, NT (60.9°N) and Yellowknife, NT (62.5°N), with an additional site east of Yellowknife along the Ingraham Trail, and two remote sites approximately 80 km northeast of Yellowknife near Lucky Lake (63.1°N) (Figure 5-1). The study transect runs southwest to northeast, perpendicular to the MAAT isotherms and the boundary between permafrost zones in the region (Figure 5-1; Heginbottom et al., 1995). To simplify, “north” and “south” are used here to represent the ends of the transect.

The regional climate is subarctic and continental, with mean annual air temperature (MAAT) for 1981-2010 of -2.5°C at Hay River and -4.3°C at Yellowknife (Figure 5-2; Environment Canada, 2019). Annual precipitation (1981-2010) is 336 mm at Hay River and 289 mm at Yellowknife, of which 41% and 55% falls as snow, respectively (Environment Canada, 2019). The climate has warmed by ~2°C over the last half century (Holloway and Lewkowicz, 2019), including by more than 1°C in the last 30 years (Figure 5-2). During the study period, however, mean air temperatures declined each year (except for Hay River in 2019), by a total of 2°C at Yellowknife (Figure 5-2). Nevertheless, the average MAAT between 2015-2019 was -1.7°C at Hay River and -3.5°C at Yellowknife, values which are still 0.8°C higher than their respective normals.

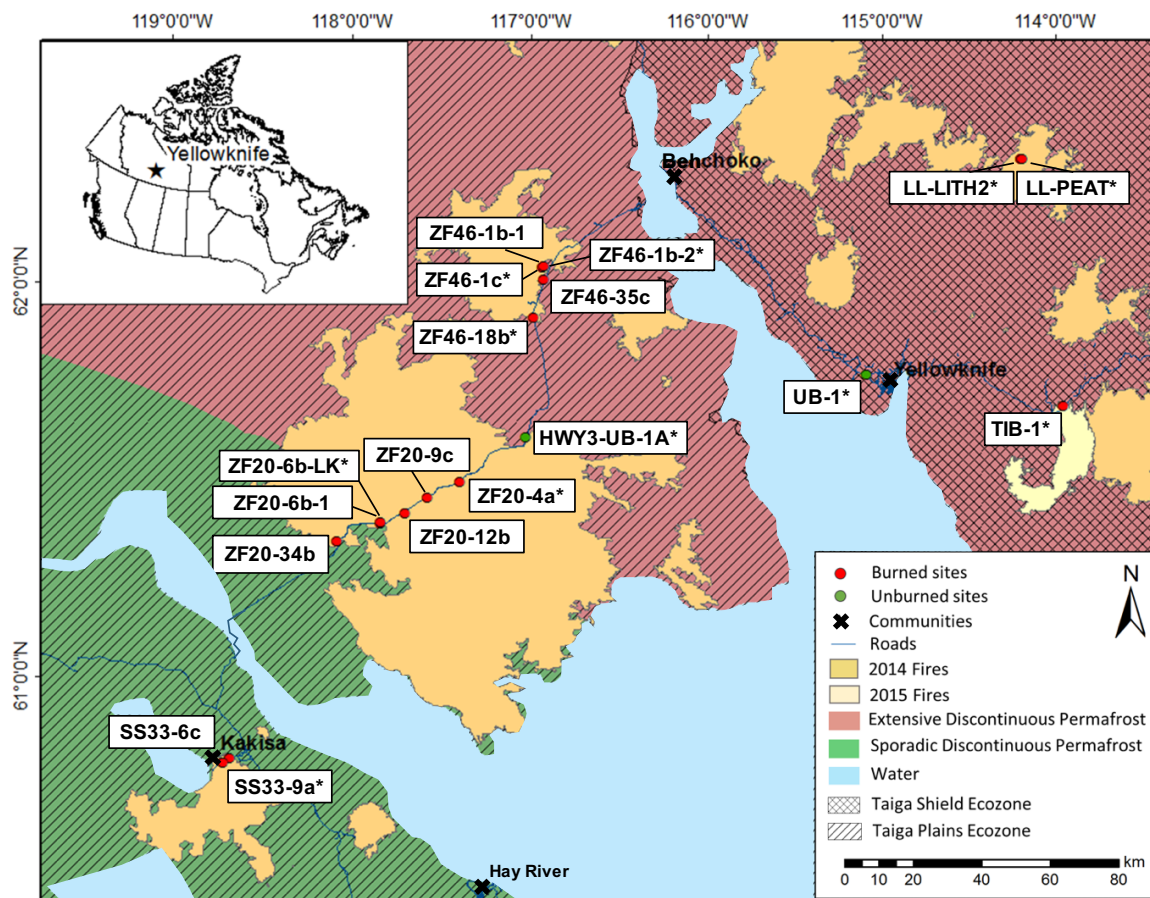


Figure 5-1: Map showing the study area in the southern Northwest Territories, including burned and unburned sites, major roads, 2014 and 2015 fires, ecozones, and permafrost zones (Heginbottom et al., 1995). Sites are labeled, and more detailed information can be found in Table C1. The 11 monitoring sites are noted with an asterisk (*).

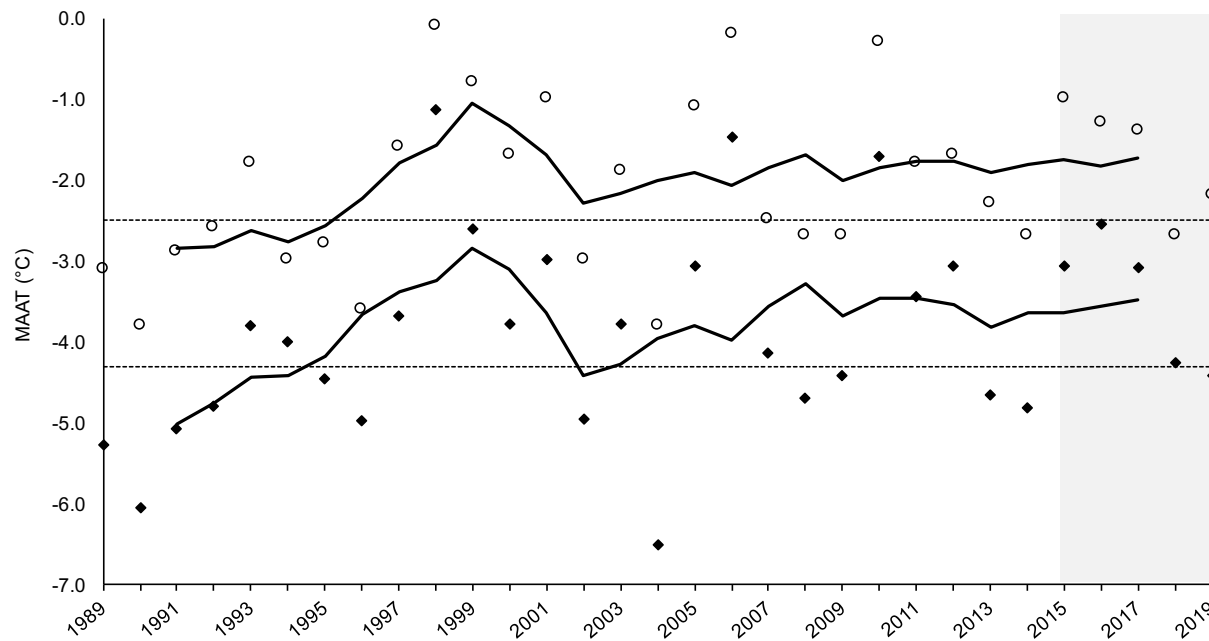


Figure 5-2: Mean annual air temperature (MAAT) at Hay River, NT (circles) and Yellowknife, NT (diamonds) from 1989 to 2019 (data source: Environment Canada, 2019). MAAT was calculated using January 1-December 31 for each year. The thick solid lines are 5-year running means. The dashed lines are the 1981-2010 climate normals. The period of study is shaded in grey. Note: some one day gaps in the record were infilled using the average of the temperatures on either side of the break.

The study transect traverses the taiga plains and the taiga shield ecozones (Figure 5-1). The former comprises flat to gently rolling terrain covered by glaciolacustrine deposits and till. The vegetation is predominantly black spruce (*Picea mariana*) forest in the lowlands and mixed-wood forest in the uplands, and the understorey is commonly composed of mosses, lichens, sedges, and Labrador tea (*Rhododendron groenlandicum*) (Ecological Stratification Working Group, 1996). The taiga shield comprises undulating exposed bedrock, with thin till and wetlands in low-lying areas. In upland areas, jack pine (*Pinus banksiana*), black spruce (*Picea mariana*) and paper birch (*Betula papyrifera*) occur on thin soils over bedrock, while black spruce (*Picea mariana*) dominates lowland areas (Ecological Stratification Working Group, 1996). Wetlands make up approximately 40% of the area, comprising peat plateaus and thermokarst bogs, with peat thicknesses of 2-6 m (Gibson et al., 2018). Peat plateaus are elevated 1-3 m above the surrounding bogs, with vegetation dominated by black spruce (*Picea mariana*), Labrador tea (*Rhododendron groenlandicum*), lichens, and *Sphagnum* (Gibson et al., 2018).

The southern part of the transect falls into the zone of sporadic discontinuous permafrost (Figure 5-1), where thicknesses are less than 30 m (Smith et al., 2015), and mean annual ground

temperatures are $> -1^{\circ}\text{C}$ (Smith et al., 2010). Near its southern margin, permafrost occurs mostly in peatlands, and has been degrading over the last century (Beilman et al., 2003; Smith et al., 2010). At the boundary with extensive discontinuous permafrost near Yellowknife, permafrost is less than 50 m thick and is generally warmer than -2°C (Wolfe et al., 2015).

5.4 Methods

5.4.1 Site Selection

Stratified random sampling was used to select areas in seven burns near the Mackenzie Highway, representing a range of vegetation types, fire severity, and fire history (including repeated burning) (Walker et al., 2018b). Further selection occurred in the field, where the first three feasibly accessible locations within each stratum were surveyed. From these random locations, sites were chosen for detailed investigation based on the presence or absence of a frost table and conditions indicative of permafrost (lowlands, black spruce canopy, wet conditions). In addition, two unburned control sites representing the two ecozones were selected along the transect. A total of 16 sites were set up in May-June 2015, with two more added in 2016 (see Table C1 for complete site information). Of the total 18 sites, seven sites were interpreted as non-permafrost and are not discussed further except for air temperature measurements (Table C1). With some exceptions, due to logistical constraints, sites were visited annually in the late summer from 2015-2019.

5.4.2 Temperature and snow depth measurements

Shielded air temperatures at a height of 1.5 m were measured at six sites using Onset Corporation (USA) Hobo Pro U23-003 loggers (accuracy $\pm 0.21^{\circ}\text{C}$; precision $\pm 0.02^{\circ}\text{C}$). Each of these represents a cluster of sites located within 20 km, an area in which air temperature was expected to vary little given the limited regional topography. All air temperature data were aggregated from 2-hour samples to daily averages using a custom R v3.3 script (Way et al., 2018). Air temperature for each site without a station was calculated as a weighted average based on distance to the two nearest stations. For site TIB-1, there was no site within 20 km to use for a weighted average, so the air temperature from site UB-1 (60 km distant) was used. Data from these sites were expected to be similar as they are at the same latitude (Figure 5-1), and data from

sites a similar distance from UB-1 was well correlated ($R^2 > 0.98$). Between-site regression analyses show that each comparison, including widely spaced sites, had an $R^2 > 0.94$ (Figure C1). To infill a gap at the beginning of the air temperature record for site HWY3-UB-1A, a linear relationship was generated from the 20 days following the missing data. The aggregated and infilled temperature record from each site was used to calculate MAAT using the hydrological year (September 1 to August 31 the following year) which was designated as the calendar year in which it ended.

Ground temperatures were measured using Onset Corporation (USA) Hobo Pro U12-008 loggers with four external thermistors (accuracy $\pm 0.25^\circ\text{C}$; precision $\pm 0.03^\circ\text{C}$). Thermistors were installed to the base of the frost table and at several depths within the seasonal active layer at all sites, and where possible, into the permafrost using the water-jet drilling technique (Table 5-1; Table C1). Ground surface temperatures were measured with Maxim Integrated™ Thermochron iButton temperature loggers (model no. DS1922L; accuracy $\pm 0.5^\circ\text{C}$) installed at 5 cm depth on the bottom of a wooden stake at each site. A second set of ground temperatures were measured in an unburned section of the site if one existed (Table 5-1). Sub-daily ground temperatures were aggregated to daily values using a custom R v3.3 script (Way et al., 2018). Gaps in the ground temperature record were infilled using a cubic spline interpolation (Riseborough and Burgess, 1996). For gaps smaller than 10 days, the 10-day period on either side of the missing data was used to generate the relationship. For gaps larger than 10 days, a 30-day period on either side was used. To infill gaps at the beginning of the ground temperature record, a linear relationship was generated from the 20 days following the missing data. Ground surface temperatures were infilled using the relationship with the air temperature at that site, where that relationship had an $R^2 > 0.8$. This was done using data for similar intervals of the year only, to avoid periods when snow was on the ground. For example, if the gap was in September, the relationship was generated using data from August 1-October 15 of each year on record. If R^2 was < 0.8 , the date ranges were adjusted due to the likelihood that snow was on the ground during that time (beginning of October or end of May). For two sites, adjusting the date ranges did not provide a sufficient R^2 value, so a linear relationship with the next closest sensor in the profile (within 10-20 cm) was used to infill gaps in the ground surface temperature record. Using the hydrological year (September 1 to August 31 the following year), the aggregated and infilled ground and surface temperature record from each site was used to calculate the mean annual ground surface

temperature (MAGST), the mean annual ground temperature at 1 m depth (MAGT), or as close as possible to this depth, and surface and thermal offsets.

Table 5-1: Depth of each sensor in boreholes at burned and unburned locations along the transect at the monitoring sites. Sites are listed from south to north, with the two unburned sites listed last. Note: GST = ground surface temperature.

Site	Instrumentation
SS33-9a	Burned = GST, 0.2, 0.4, 0.6, 0.8, 1, 1.5, 2.5, 3.1 m Unburned = GST, 0.1, 0.3, 0.5, 0.7, 1, 1.5, 2, 2.4 m
ZF20-6b-LK	Burned = GST, 0.3, 0.5, 0.8, 1, 1.5, 2, 2.5, 3, 3.2 m
ZF20-4a	Burned = GST, 0.2, 0.4, 0.5, 0.6, 0.8 m Unburned = GST, 0.2, 0.4, 0.6, 0.8 m
ZF46-18b	Burned = GST, 0.2, 0.3, 0.5, 0.6 m
ZF46-1b-2	Burned = GST, 0.3, 0.5, 1, 1.5, 1.9 m Unburned = GST, 0.5, 0.8, 1, 1.5 m
ZF46-1c	Burned = GST, 0.2, 0.4, 0.6, 0.8 m; Unburned = GST, 0.1, 0.2, 0.4, 0.6 m
TIB-1	Burned = GST, 0.2, 0.5, 1 m
LL-PEAT	Burned = GST, 0.8, 1.5, 2.2 m
LL-LITH2	Burned = GST, 0.8, 1, 1.5, 2, 2.2, 3, 4, 5, 6, 7, 8 m
HWY3-UB-1A	Unburned = GST, 0.3, 0.5, 1, 1.5 m
UB-1	Unburned = GST, 0.5, 1, 1.5, 2, 3, 4, 6, 7.4 m

At each station, snow depth was approximated using data collected continuously by Thermochron iButtons placed at various intervals on a wooden stake (typically 10 cm, 20 cm, 30 cm, 40 cm, 50 cm, 60 cm, 80 cm) (Lewkowicz, 2008). Snow depth was estimated by comparing temperatures at the different heights since iButtons covered by snow experience smaller diurnal temperature variations than those exposed in the air (Lewkowicz, 2008). Raw iButton data were first reformatted (Way et al., 2018) and then used to calculate daily temperature ranges at each height. Snow depths were inferred using the absolute differences between the 2-day future running mean and standard deviation between the highest depth (typically 80 cm) and all other depths on the iButton stake. Snow was deemed to be present at a certain height if:

- the temperature value recorded was below a set temperature threshold;
- the maximum temperature value recorded in the subsequent 24 hours was below the temperature threshold;
- the difference in 2-day future running means was below the threshold; and,
- the difference in the 2-day standard deviation was below the set threshold.

A threshold of 1 was used for both temperature and standard deviation, as this value best reduced noise in the data. If snow was present at one height, and not at the height above, snow depth was assumed to be mid-way between the iButtons. Manual smoothing was required to remove outliers where the threshold was too low or too high and to avoid missing the presence of snow cover. Corrected daily snow depths were used to derive maximum snow depth and snow-depth days (SDD) as m.d. Due to the vertical spacing of iButtons, there could be up to 10-20 cm error in snow depth values, and these uncertainties are exacerbated by iButton failures, which result in larger gaps between sensors. Due to sensor failure, it was not possible to calculate snow depth for site HWY3-UB-1A for 2017.

Annual freezing and thawing degree-day (FDD and TDD) totals were calculated from the air and surface temperature records by summing daily mean temperatures below and above 0°C, respectively. For this calculation, the freezing season was assumed to be October 1 to April 30 of the next calendar year, and the thawing season from May 1 to September 30 of the same year (Smith et al., 2015). Degree-day totals were used to calculate n-factors, which are empirical transfer functions that account for the local influence of snow cover, vegetation and other surface characteristics on ground temperatures (Morse et al., 2012; Smith and Riseborough, 2002; Throop et al., 2012). Surface thawing and freezing n-factors are defined as:

$$n_T = \frac{TDD_s}{TDD_a}; n_F = \frac{FDD_s}{FDD_a}$$

where n_T is the thawing n-factor, TDD_s is the surface thawing degree-day total, TDD_a is the air thawing degree-day total, n_F is the freezing n-factor, FDD_s is the surface freezing degree-day total, and FDD_a is the air freezing degree-day total. In this analysis, n-factors are used as indicators of changes to heat transfer and surface conditions following fire. TDD and n_T could not be calculated for 2015 because instruments were installed after the start of the thaw season, nor for 2019 because records ended in September.

Finally, surface, nival, and thermal offsets were calculated for each site, using the equations:

$$\text{Surface offset} = \text{MAGST} - \text{MAAT}$$

$$\text{Nival offset} = \text{MGST}_F - \text{MAT}_F$$

and

$$\text{Thermal offset} = \text{TTOP} - \text{MAGST}$$

where MGST_F and MAT_F are the mean ground surface and mean air temperatures measured during the freezing season, and TTOP is the mean annual temperature at the top of permafrost (we used the MAGT at 1 m depth) (Smith and Riseborough, 2002). The freezing season was defined as starting on the first day the mean daily ground surface temperature dropped and remained below 0°C, and ending on the day it remained above 0°C.

5.4.3 Direct current electrical resistivity tomography surveys

An ABEM LS multi-electrode, two-dimensional system was used to conduct electrical resistivity tomography (ERT) surveys at all sites, to investigate permafrost distribution and estimate its thickness, and to monitor changes over time. Stainless steel electrodes were arranged in a Wenner array, with a minimum spacing of 2 m and a depth of investigation of 24 m for 160 m-long surveys. The electrodes were left in place throughout the study to ensure that there were no changes in their positions. Measurements were made by injecting current into the ground through two current electrodes, and measuring the resulting voltage difference at two potential electrodes, then calculating an apparent resistivity (p_a) value from:

$$p_a = k V / I$$

where V is the voltage, I is the current, and k is the geometric factor which depends on the electrode array (Loke, 2001). The apparent resistivity is the value of an electrically homogenous subsurface that results in the measured relationship between the current and potential electrodes for a certain array and spacing (Loke, 2001). To determine the true resistivity of the heterogeneous ground, tomograms were produced using the robust inversion method in the software package RES2DINV, which performs smoothness-constrained inversions of measured apparent resistivity using finite difference forward modelling (Loke, 2001). Iteration was carried out until the RMS error fell below 5%, or to the fifth iteration, whichever occurred first. Model sensitivity was calculated in RES2DINV (Loke, 2003), and model blocks with sensitivity less than 0.1 were faded in the tomograms (Way et al., 2018). All profiles were topographically

corrected before inversion using elevations obtained in the field with a ZipLevel Pro-2000 altimeter (2 mm resolution). Where possible, the thresholds between frozen and unfrozen soil predicted by Holloway and Lewkowicz (2019) was used to interpret ERT surveys. To calculate changes in resistivity over time, the time-lapse technique was employed in RES2DINV, where the inversion of the first year's resistivity data was used as a reference model to constrain the data collected on subsequent dates. This method produces a profile that shows the percentage variation in resistivity between the two dates while reducing noise-related time-lapse artefacts (Loke, 2001; Miller et al., 2008; Oldenborger et al., 2007).

5.4.4 Frost table depths

Frost table measurements were used to monitor near-surface changes to the permafrost and to validate ERT surveys. A 1 cm diameter titanium probe was pushed into the ground at each electrode until it met refusal (Brown et al., 2000; Mackay, 1977). Frost table depths were measured between August 29 and September 15, with the exception of 2015, when measurements were taken in mid-August. The 2015 measurements were adjusted by 6-11% according to the square root of additional TDD_a of each site between the day of measurement in 2015 and the mean date of measurement in 2016-2018. Since no air temperature was recorded at the sites prior to June 2015, the air temperature at Yellowknife was used to calculate the TDD for that year (Environment Canada, 2019). The relationship between the Yellowknife air temperature and the temperature of the measured period at each site was used to adjust the Yellowknife TDD ($R^2 > 0.95$).

No frost table was encountered within 110 cm at many probing spots, a number which grew in later years, which meant that permafrost was absent or that the frost table depth exceeded 110 cm. To facilitate inter-year comparisons, the average of the first quartile (Q1) of thaw depths are presented, and values greater than 110 cm are included in the sample as equal to 110 cm. Since maximum thaw depth in the area occurs after mid-September (Holloway and Lewkowicz, 2020; Figure C2), these depths represent a minimum active layer thickness.

5.4.5 Ground subsidence

Ground subsidence was examined annually at all sites from 2016 onwards. A Trimble R7 dGPS (cm-accuracy) was used to measure the elevation at the start of the ERT transect, and a

ZipLevel Pro-2000 altimeter (2 mm resolution) measured the relative elevation of each electrode. However, despite being left in place for one hour with sampling every 15 s and post-processing of the data with the Precise Point Positioning tool (Natural Resources Canada, 2020), the average standard deviation of the dGPS positions (which incorporates instrument error) was 10 cm. This error is greater than the relative elevation change registered at the majority of sites, rendering the data unusable. Instead, ZipLevel surveys at eight of the sites were used to calculate the change in ground surface levels relative to anchor points (e.g. bedrock) or changes were compared to unburned sections of the profile. It was necessary to manually adjust some profiles to account for issues when measuring pond water level.

5.4.6 Burn severity measurements

Burn severity was estimated at each electrode using a descriptive index with a value of 0-3 based on the condition of the surface organic mat and canopy combustion (Table 5-2). These values were summed across the transect to obtain a site-wide “Burn Severity Index”. In addition, adventitious roots were measured along the transect, where possible, to determine the depth of burn. These roots occur on black spruce trees as a response to the pre-burn organic layer accumulating upwards around the stem of the tree. After the organic layer is consumed during fire, the nubs of adventitious roots remain and their heights above the ground surface provide an estimate of the thickness of the organic mat consumed during the fire (Boby et al., 2010; Kasischke and Johnstone, 2005). Three adventitious root measurements were made at each electrode where a black spruce tree was within 2 m. Adventitious root heights were calibrated using the residual organic layer depth and the mean depth of the adventitious roots measured in unburned stands at the same sites (Walker et al., 2018a). Lastly, pits were dug at each site to measure the residual organic layer depth.

Table 5-2: Descriptive index of burn severity.

Burn Severity	Qualitative Description
Severe burn (3)	Organic mat completely burned exposing mineral soil; all trees have fallen
Moderate burn (2)	Patches of organic material are intact but are charred and patches of mineral soil are exposed; trees remain standing
Slight burn (1)	Organic material remains intact but is charred
Unburned (0)	Unburned patch along the transect

5.4.7 Soil sampling and laboratory measurements

Soil samples from the active layer and upper permafrost were collected at some sites (Table C1) in 2016 using a motorized SIPRE corer with a 75 mm-diameter core barrel. Where coarse surficial material made coring impossible, pits were dug and soil samples taken from the active layer. Core samples were shipped frozen to Ottawa for laboratory analysis, including excess ice content, bulk density, gravimetric moisture content (GMC), grain size, and organic matter content using the loss on ignition (LOI) technique (Heiri et al., 2001).

Excess ice is the volume of supernatant water present when a column of permafrost is thawed (French, 2018). Permafrost samples were thawed, poured into glass beakers with straight sides, weighed, and allowed to settle for 24 hours. Excess ice (E_i) was calculated using the following formula:

$$E_i(\%) = \left[\frac{W_v \times 1.09}{S_v + (W_v \times 1.09)} \right] \times 100$$

where W_v is the volume of supernatant water (cm^3), multiplied by 1.09 to estimate the equivalent volume of ice, and S_v is the volume of saturated sediment (cm^3) (Kokelj and Burn, 2003). The volumes of supernatant water and sediment were measured, and then samples were dried in an oven at 105 °C for 24 hours before the dry weight was determined. Bulk density was calculated by dividing the dry weight of the core by the total volume (sediment and water). The GMC was calculated using:

$$GMC(\%) = \left[\frac{W_w - W_d}{W_d} \right] \times 100$$

where W_w is the wet weight (g) and W_d is the dry weight (g) (van Everdingen, 2005).

Grain size distributions were determined using a Microtrac S3500 laser particle size analyzer (Microtrac Inc., PA) (LPSA), which measures particles from 0.02 to 2800 µm in diameter. Organic matter was removed from the fine fraction using 30% peroxide (H₂O₂), following methods by Verret et al. (2017). The remaining mineral soil sample was dispersed using a 5% hexametaphosphate solution before a sample aliquot was introduced into the LPSA chamber. Replicates were analyzed for 5% of the samples to ensure the precision of the instrument, which was within 2-4%. However, error for samples containing sand was substantially higher (between 9-274%), due to anomalously large grains causing bias. To obtain a more accurate average grain size distribution, such samples were analyzed five times each. Grain size was not conducted on peat samples because the LPSA does not differentiate between sediment and sediment-sized organic matter. Before analysis, the samples were passed through a 2 mm sieve to remove the coarse fraction. Only three sites had particles >2 mm, and at two of these (ZF46-1c and ZF46-18b), gravel made up less than 10%. The remaining site (ZF20-4a) contained between 36-86% gravel in the soil profile. Since gravel composed a significant portion at only one site, removing the coarse fraction of samples was considered acceptable, particularly since it is the fine fraction that is of interest to determine frost susceptibility (French, 2018). The final result is the grain size distribution for each sample, with an average grain size in µm.

LOI was completed by placing a subsample in a crucible positioned into a LECO thermo-gravimetric analyzer with an integrated balance (model TGA701). The analyzer first reached a temperature of 105°C, weighed the sample when the variation in weight reached 0.1% (LOI₁₀₅), then the temperature increased to 550°C to combust the organic matter content, and then weighed the sample again (LOI₅₅₀). The organic matter content was calculated as follows:

$$\text{Organic matter content (\%)} = (LOI_{105} - \frac{LOI_{550}}{LOI_{105}}) \times 100$$

(Heiri et al., 2001). The accuracy of the analysis was determined from a calcium oxalate standard, with an analytical error of ±0.13%.

5.5 Results

Observations at the permafrost monitoring sites between 2015-2019 are described here. Variables analyzed but which showed no temporal trends are shown in Appendix C. Note that statistical analysis was not undertaken when variables changed through time due to the lack of

independence. As such, associations are described qualitatively rather than in terms of statistical significance.

5.5.1 Post-fire changes in air and near-surface temperatures

The four-year monitoring period coincided with a return to average temperatures, although the running mean remained well above the long-term normals at both Yellowknife and Hay River (Figure 5-2). MAAT was highest in 2016, and median values at both burned and unburned monitoring sites subsequently fell by about 2°C over the measurement period, but with the least amount of change between 2018 and 2019 (Figure 5-3). Median MAAT values were 0.1-0.4°C warmer at the unburned sites (n = 1-2) than at the burned ones (n=8-9), although these values partially fall within the analytical precision of the sensor ($\pm 0.2^\circ\text{C}$). However, all unburned values fell within the interquartile range (IQR) of burned sites in all four years (Figure 5-3).

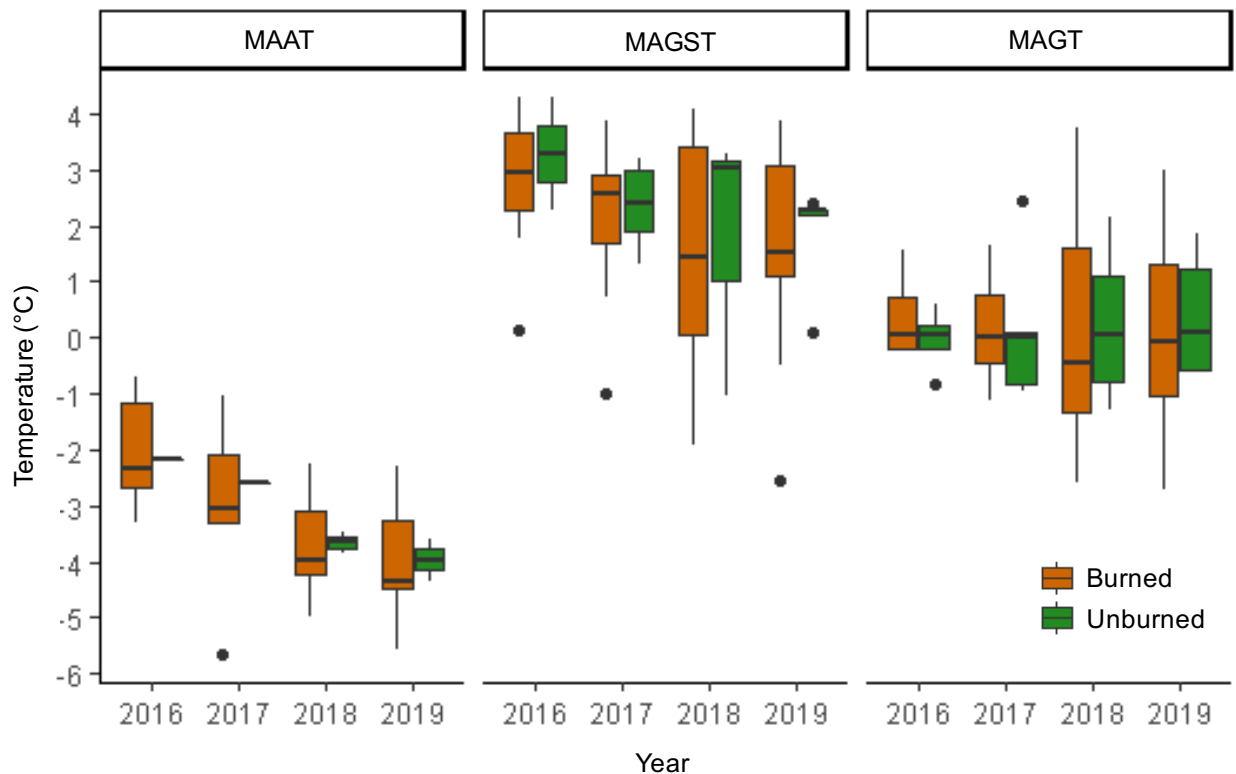


Figure 5-3: Boxplots of mean annual air temperature (MAAT), mean annual ground surface temperature (MAGST), and mean annual ground temperature at 1 m depth (MAGT) for all burned and unburned sites, as well as unburned sections of burned sites, for hydrologic years 2016-2019. For MAAT, burned n = 8-9, and unburned n = 1-2. For MAGST, burned n = 6-9, and unburned n = 2-6. For MAGT, burned n = 5-8, and unburned n = 4-6. Boxes extend from lower hinge (first quartile) to the upper hinge (third quartile) with the solid line as the median. Whiskers extend to the largest or smallest value no further than $1.5 \times \text{IQR}$ from the hinge (where IQR is the inter-quartile range; Tukey, 1977). Dots beyond the whiskers are outliers.

Median values for MAGST consistently exceeded 0°C at all sites (Figure 5-3), but also showed a cooling trend, declining at burned sites by 1°C from 2016-2018, and remaining stable between 2018 and 2019 (Figure 5-3). There were three sites with sub-zero MAGST, including one of the unburned sites (HWY3-UB-1A in 2018 only) and the two northernmost burned sites at Lucky Lake (for three years at one lithalsa and two years at the other) (Table C2). There was a high degree of overlap between burned and unburned sites, with median values being higher by 0.4-1.6°C at unburned sites, except in 2017. The range (excluding outliers) increased at both burned and unburned sites between 2017 and 2018 by 2.9°C and 2.4°C respectively, including to 6°C for burned locations in 2018, which was more than twice that of MAAT in the same year (Figure 5-3). This occurred mostly at the colder end of the distribution, likely due to the progressive decrease in MAAT over the same period.

Median MAGT values at 1 m depth at both the burned and unburned sites were all within ± 0.5 of 0°C and showed no trend through time (Figure 5-3). Despite this consistency, however, ranges increased in both groups. The range (excluding outliers) of MAGT at burned sites was 1.8-6.3°C between 2016 and 2019, and as with MAGST, was greatest in 2018. However, unlike MAGST, the range increased at both the warmer and colder ends of the distribution, with the maximum temperature reaching 3.7°C in 2018. This was partly due to inconsistencies in sample site numbers between years (Figure 5-3) due to additional burned and unburned sites being monitored after 2016 (Table C1) and equipment failures at other sites. MAGT were generally warmer in coarse textured soils and those with low organic contents (Figure 5-4a and b). There was an inverse relationship between moisture content and MAGT, and the active layer was wetter at unburned sites than burned sites (Figure 5-4c). Most sites showed little change between years, with the exception of ZF20-6b-LK which warmed by 2.8°C between 2016 and 2019 (Figure 5-4a).

Typically, there were about 1000 more degree days at the surface in the thawing season than in the freezing season, at both burned and unburned sites (Figure 5-5). However, the inter-site range in FDD_s was large, often exceeding 1000 degree days, and there were eight outliers (Figure 5-5). For both burned and unburned sites, FDD_s were stable from 2016-2018, with median values within about 100 degree days. In 2019, the median FDD_s decreased by 194 and 167 degree days at burned and unburned sites respectively. For TDD_s, median values at burned

sites were within 100 degree days in all years, while at unburned sites the median dropped by nearly 300 degree days in 2018. There was extensive overlap between burned and unburned values for both FDD_s and TDD_s (Figure 5-5).

5.5.2 Post-fire changes in frost table depths

Frost tables, measured as the mean of the first quartile of the distribution, were generally deeper at burned sites than unburned sites and they changed more over the 4-year period (Figure 5-6). Thaw depths increased at all sites between 2015 and 2019, but most of this change occurred in the first two years, and there was a slight decrease at the two unburned sites and one of the burned ones between 2018 and 2019. The greatest change in frost table depths was recorded at site ZF46-1b-2, where the average Q1 of frost table depths increased from 68 cm to >110 cm between 2016 and 2017. Sites that experienced the most change in frost table depths had coarse textured soil (Figure 5-6a), low organic content (Figure 5-6b), and low gravimetric moisture content (Figure 5-6c).

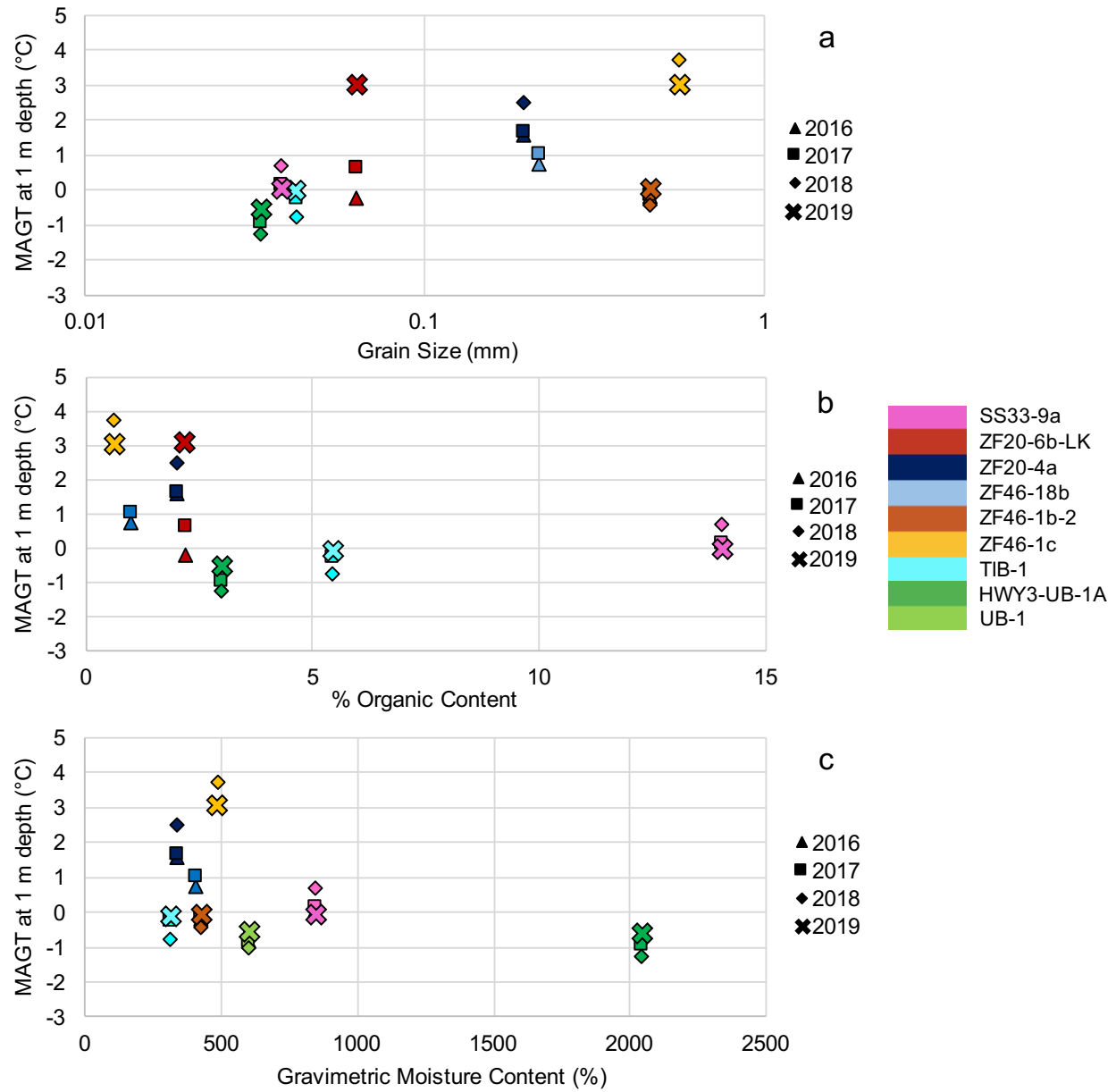


Figure 5-4: Changes in mean annual ground temperature (MAGT) at 1 m depth at the monitoring sites in relation to: (a) average grain size in the soil column (n=8), (b) average organic content (n=7), and (c) average gravimetric moisture content of organic soils in the active layer (n=8). Note: a single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green. The full range of soil characteristics can be found in Table C3.

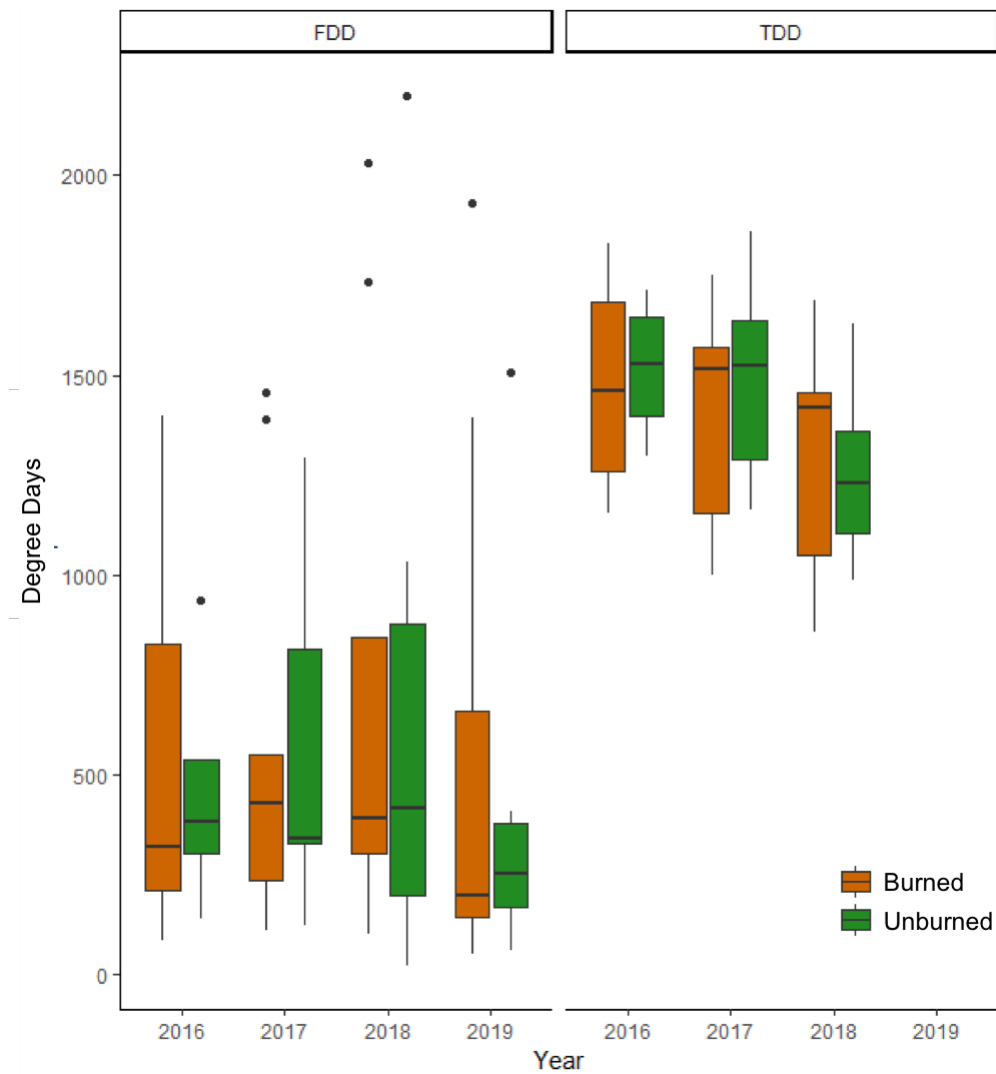


Figure 5-5: Ground surface thawing degree days (TDDs) and freezing degree days (FDDs) for all burned and unburned sites, as well as unburned sections of burned sites, between 2016 and 2019. TDDs values were not calculated for 2019 as the measurement record stopped before the end of the thaw season. For FDDs and TDDs, burned $n = 7-9$ and unburned $n = 4-6$. Boxes extend from lower hinge (first quartile) to the upper hinge (third quartile) with the solid line as the median. Whiskers extend to the largest or smallest value no further than $1.5 * \text{IQR}$ from the hinge (where IQR is the inter-quartile range; Tukey, 1977). Dots beyond the whiskers are outliers.

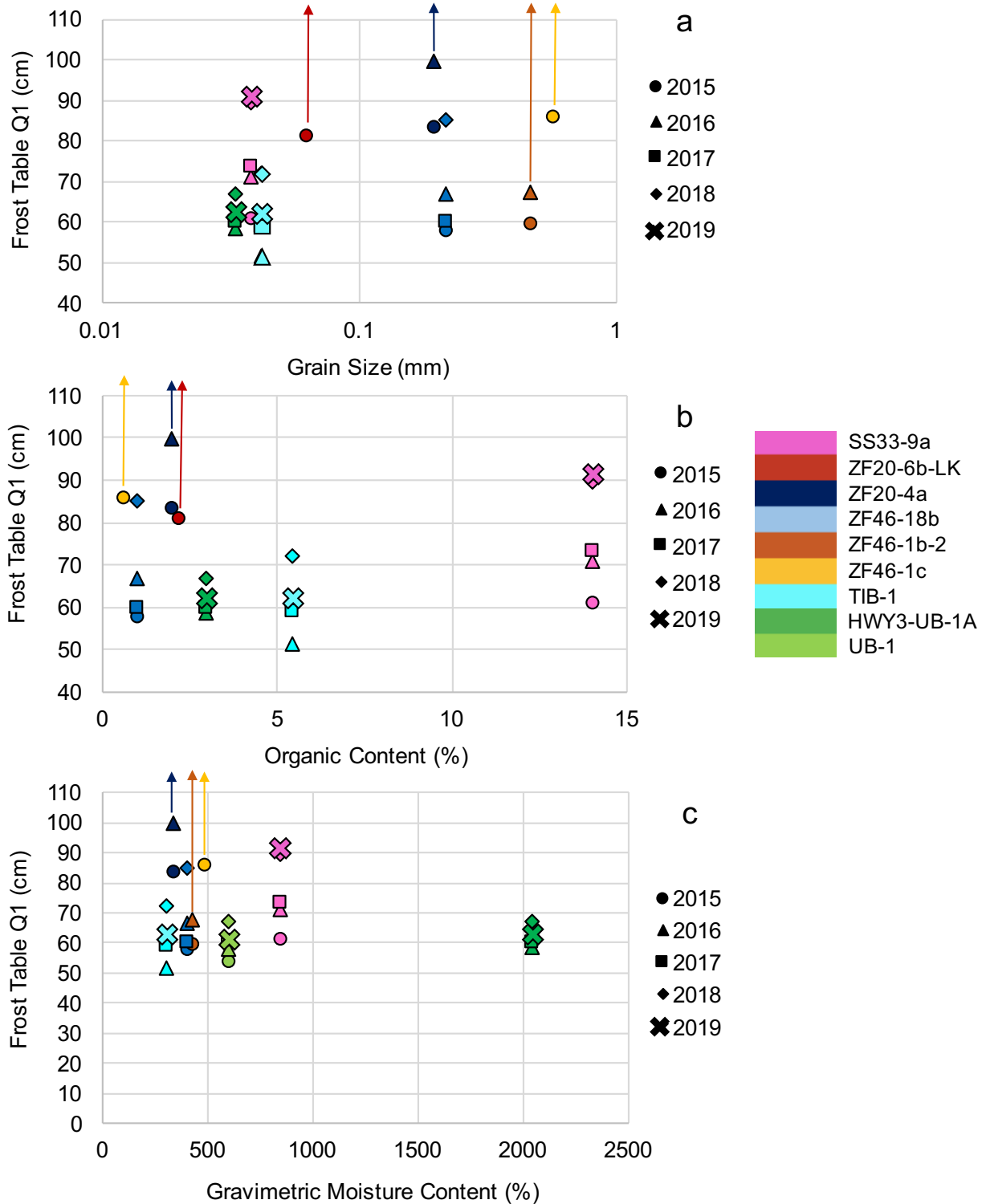


Figure 5-6: Average first quartile (Q1) frost table depths for each site and years based on: (a) the average grain size in the soil column (n=8), (b) the average organic content (n=7), and (c) the average gravimetric moisture content of organic soils in the active layer (n=8) at the permafrost monitoring sites. Note: arrows indicate that measurements in later years were greater than the length of the probe; a single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green. The full range of soil characteristics can be found in Table C3.

The pattern of greater increases in frost table depths over time at burned sites was consistent within individual sites (Figure 5-7). Overall, frost tables were deeper and increased more at burned sites (Figure 5-7b and d) compared to unburned sites (Figure 5-7a and c). On average, the Q1 of frost table depths increased by 47% and 24% at burned sites ZF20-4a and TIB-1, respectively, and by only 10% at the two unburned sites. At site ZF20-4a, there was no frost table within the top 110 cm in 2018 (Figure 5-7d).

5.5.3 Post-fire changes in resistivity

Average apparent resistivity at the monitoring sites ranged from 38 to $> 3000 \Omega\text{m}$ (Figure 5-8). Values declined between 2015 and 2018 at all sites, including the unburned ones. The greatest change in average resistivity over the period of record was at sites with organic layers less than 40 cm thick. For example, resistivity decreased by more than 60% between 2015 and 2018 at sites ZF20-6b-LK and ZF46-1c. Sites that experienced the most relative change over time also had the lowest average resistivities.

Site tomograms also show a decrease in resistivity in the near surface between 2015 and 2018, but to a greater degree and depth at the burned sites (Figure 5-9b and d). This accords with the observed increases in frost table depths at the same sites (Figure 5-7). For most sites, resistivities at depths greater than 4 m are within $\pm 10\%$ of the values obtained in 2015 (or 2016). The exception is ZF20-4a, which showed decreases in resistivity across most of the profile of $> 70\%$ to depths of at least 10 m (Figure 5-9d). However, there were also increases in resistivity in small parts of all the profiles, particularly for ZF20-4a, which included a set of model blocks whose resistivities increased by $> 70\%$. These are probably artefacts due to modelling the rapid changes in resistivity with depth at the interface between thawed and frozen soil as there is no obvious physical reason for such an increase.

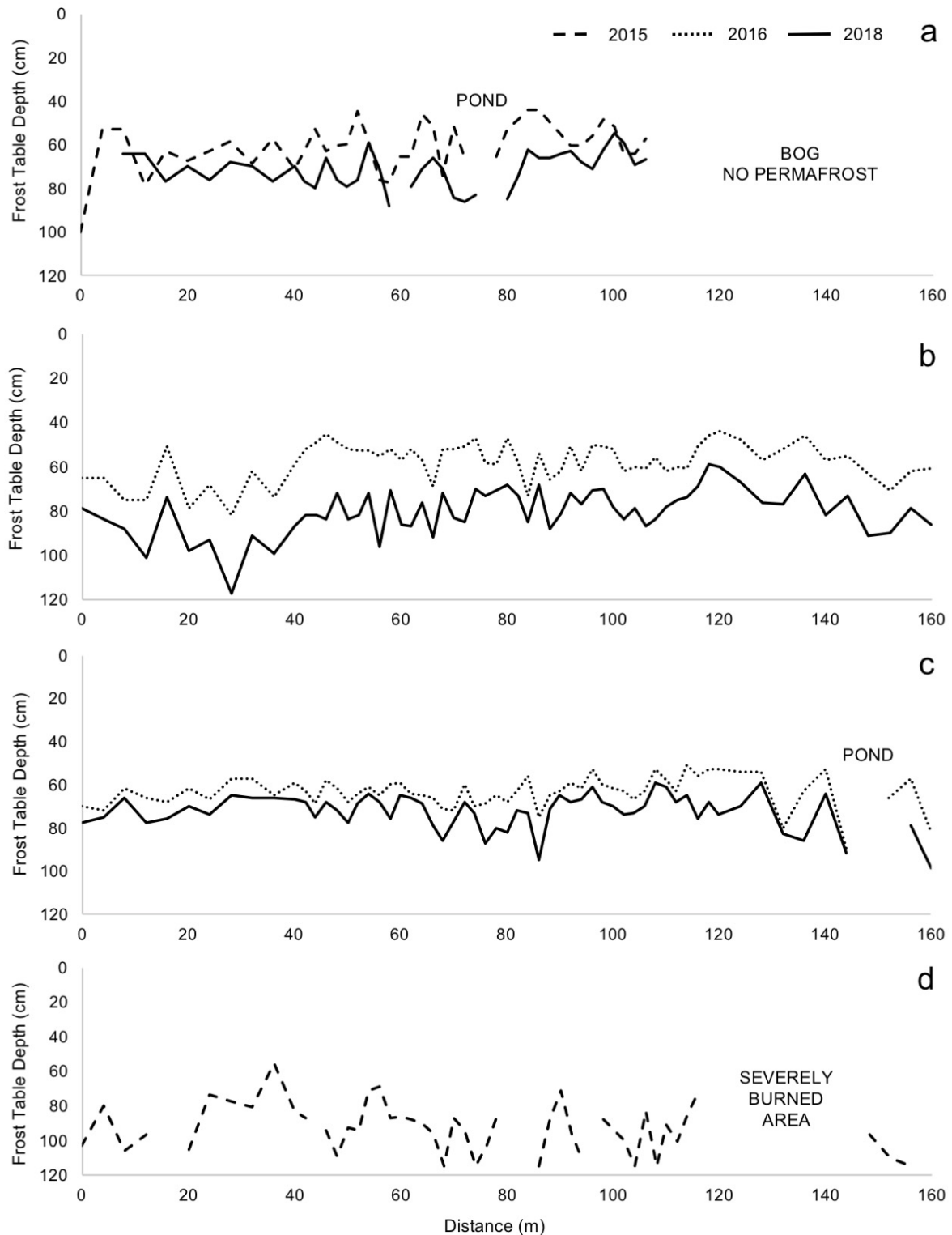


Figure 5-7: Comparison between frost table depths measured in 2015 or 2016 and 2018 at site (a) UB-1, (b) TIB-1, (c) HWY3-UB-1A and (d) ZF20-4a. Frost tables were measured at 2-4 m intervals along transects at the sites. Blank areas indicate no frost table within 110 cm of the surface. Note for ZF20-4a all measurements in 2018 were greater than 110 cm. 2015 measurements were adjusted according to the square root of additional thawing degree days between the day of the survey and the comparable date in 2018.

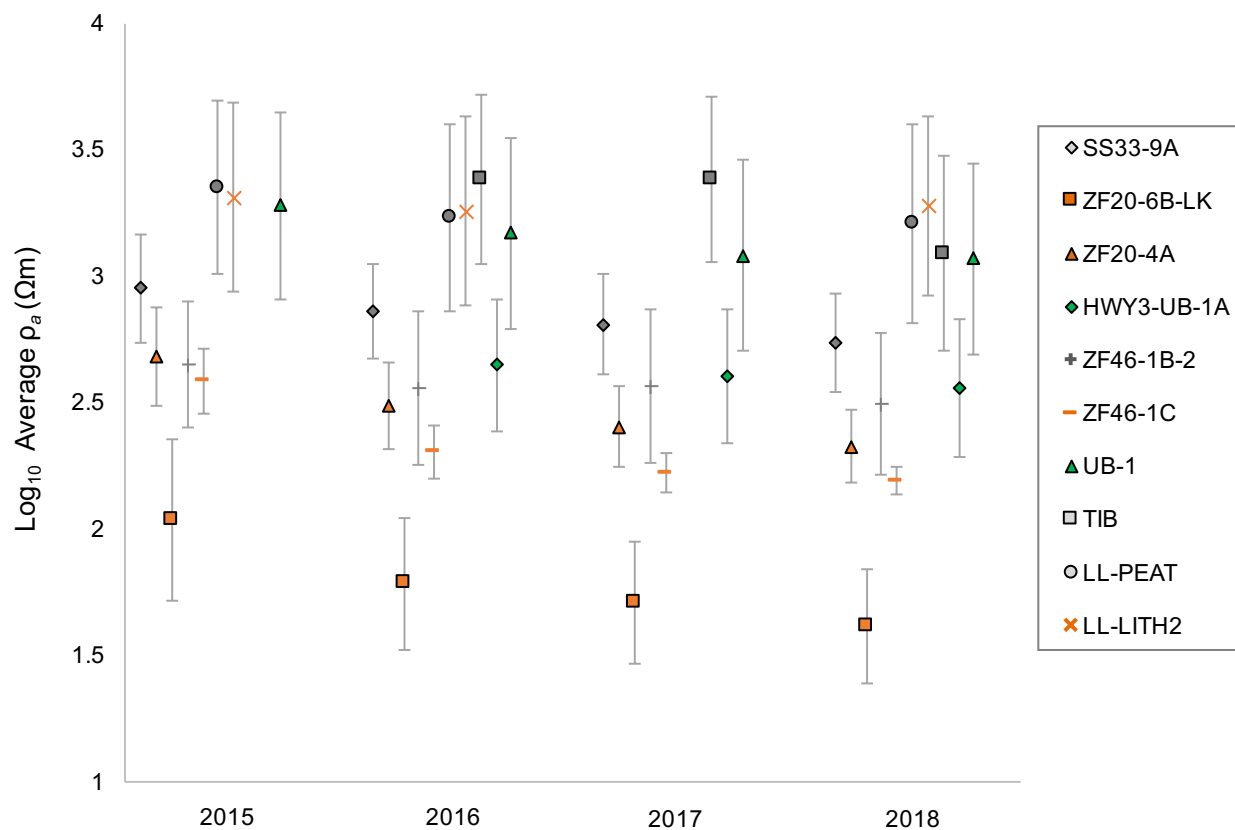


Figure 5-8: The decadic logarithm ($\log_{10}$) of the average apparent resistivity (ρ_a) for permafrost monitoring sites between 2015 and 2018. Note the year has been offset by one for each site to allow for visualization. Sites in the legend are organized from south (top) to north (bottom) along the study transect. Orange symbols represent burned sites with thin organic layers (<40 cm), grey symbols represent burned sites with thick organic layers (>40 cm), and green symbols represent unburned sites. Bars show the standard deviation.

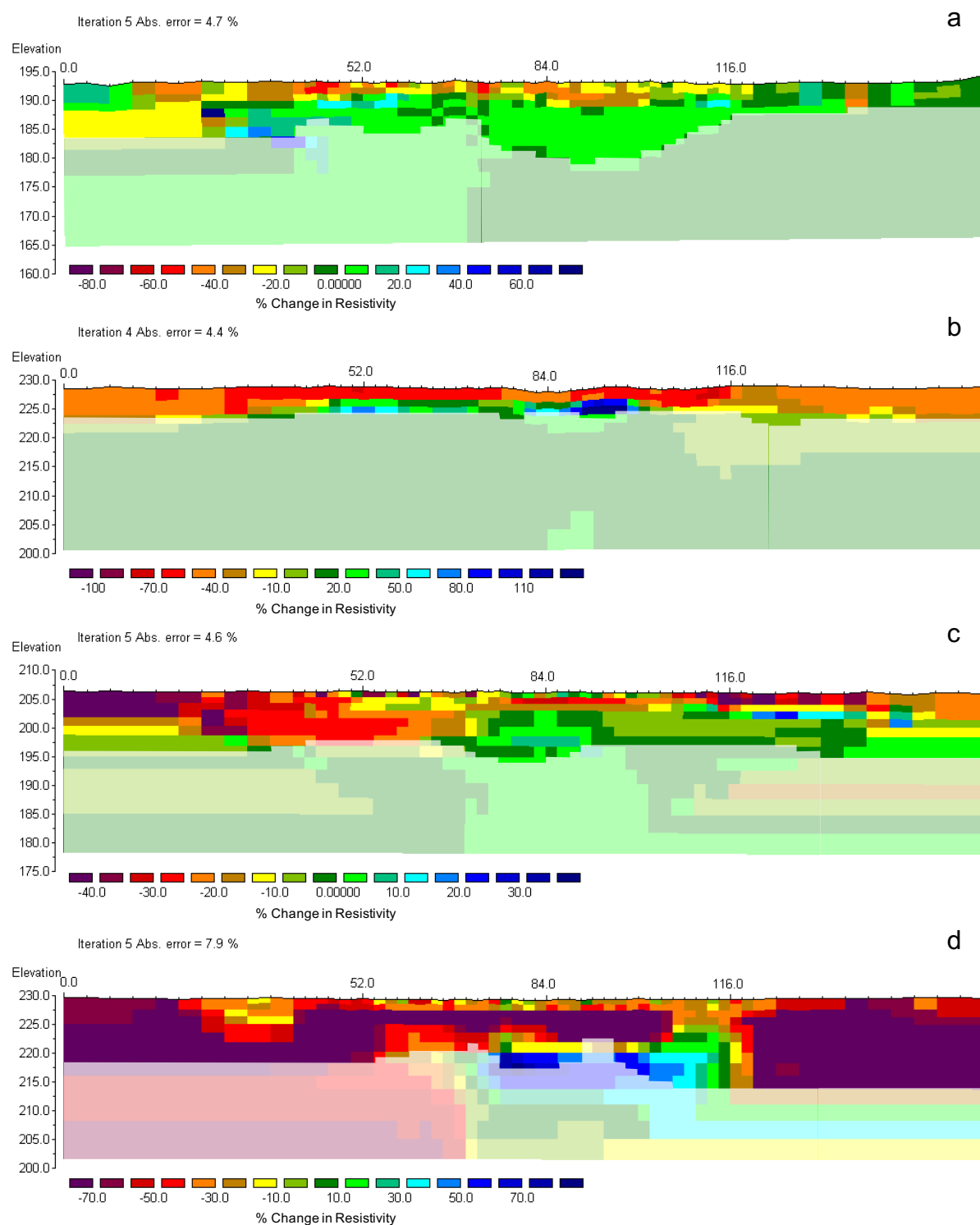


Figure 5-9: Time-lapse electrical resistivity profiles (160 m long) presented as the percent (%) change in resistivity between 2015/16 and 2018 for (a) an unburned site (UB-1) and (b) a burned site (TIB-1) in the taiga shield ecozone, and (c) an unburned site (HWY3-UB-1A) and (d) a burned site (ZF20-4a) in the taiga plains ecozone. Note that the range of resistivity change exceeds the scale end-points, but these cut offs were chosen to assist in visualization. Areas where resistivity measurements had sensitivity values below 0.1 have been faded out. Time-lapse profiles from the remaining sites are shown in Figure C6.

Comparisons of resistivity were made for model blocks at 3.3-4.6 m depth (mid-point of the model block is ~4 m) in order to focus on areas below the active layer that could be impacted by talik development or internal thaw. Resistivity values decreased at all sites between 2015 and 2018, and slightly more at burned sites (43% decrease, n=9) than unburned sites (36% decrease, n=2) (Figure 5-10). However, this pattern is site-specific, as some burned sites showed little change, and one of the unburned sites (UB-1) showed a large decrease in resistivity between 2015 and 2016. Overall, the greatest decreases in resistivity were recorded between 2015 and 2016. The greatest change in resistivity occurred at site ZF20-6b-LK, where the average resistivity fell from 114 Ωm to 29 Ωm between 2015 and 2018. At TIB-1, resistivity dropped substantially between 2017 and 2018 (Figure 5-10). Sites that experienced the most change over time had low bulk densities, and sites with low bulk density had higher resistivities (Figure 5-10).

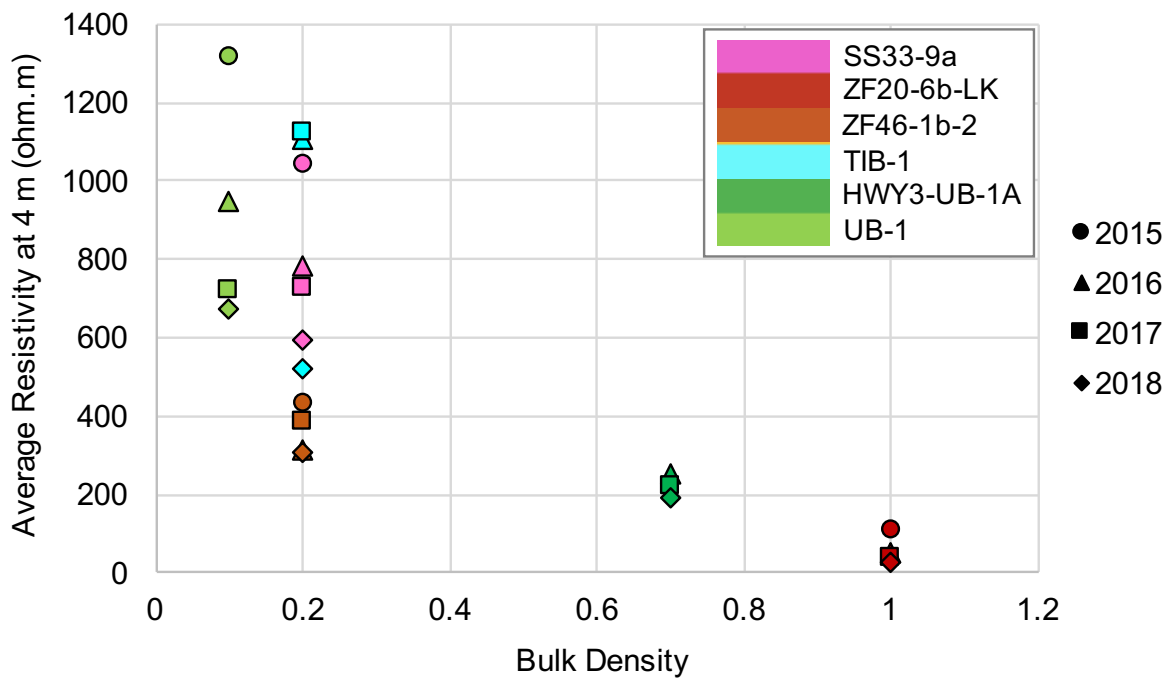


Figure 5-10: Average resistivity at 4 m depth in individual years compared to the average bulk density. Note: linear y-axis scale. A single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green. The full range of soil characteristics can be found in Table C3.

5.5.4 Impact of burn severity on permafrost

Fire affected the sites differently depending on their pre-existing characteristics. As might be expected, the residual organic layer was thicker and less of the total organic layer was burned at sites with high gravimetric moisture content, although the latter relationship was not significant (Figure 5-11). In turn, these characteristics affected the ground thermal regime. For example, sites which showed the greatest change in surface temperature over time had generally lost more of the organic layer (Figure 5-12a), although the site with the highest portion of organic layer loss did not show the greatest change over time. A clearer trend emerged, however, in terms of the impact of organic layer loss on frost table depths (Figure 5-12b). Frost tables were deeper overall at burned sites, and more change occurred over time at sites with a greater portion of organic layer loss (Figure 5-12b). There was also a positive relationship between residual organic layer depth and resistivity at 4 m depth (Figure 5-12c).

Burn severity was often non-uniform within sites resulting in variable impacts on permafrost. For example, severely burned areas were present in parts of site ZF20-4a (Figure 5-13a and f) while other parts of the transect were unburned (Figure 5-13a and e). In 2015, the frost table was 88 cm on average across the transect, with some areas greater than the 110 cm probe and some refusal by clasts, but by 2018 there was no measurable frost table within 110 cm of the surface, including an unburned section of the transect (Figure 5-13b). The MAGT at 1 m depth in the burned borehole was more than 2°C warmer than in the unburned borehole ~100 m away (Table C2). The ERT profile from 2015 shows 10-15 m of permafrost, with a small through-going talik between 108-118 m along the transect. The repeat survey in 2018 shows the highest measured resistivities in an unburned area of the transect, deeper frost tables in burned areas, particularly in the severely burned section between 120-160 m, and expansion of the talik (Figure 5-13c and d).

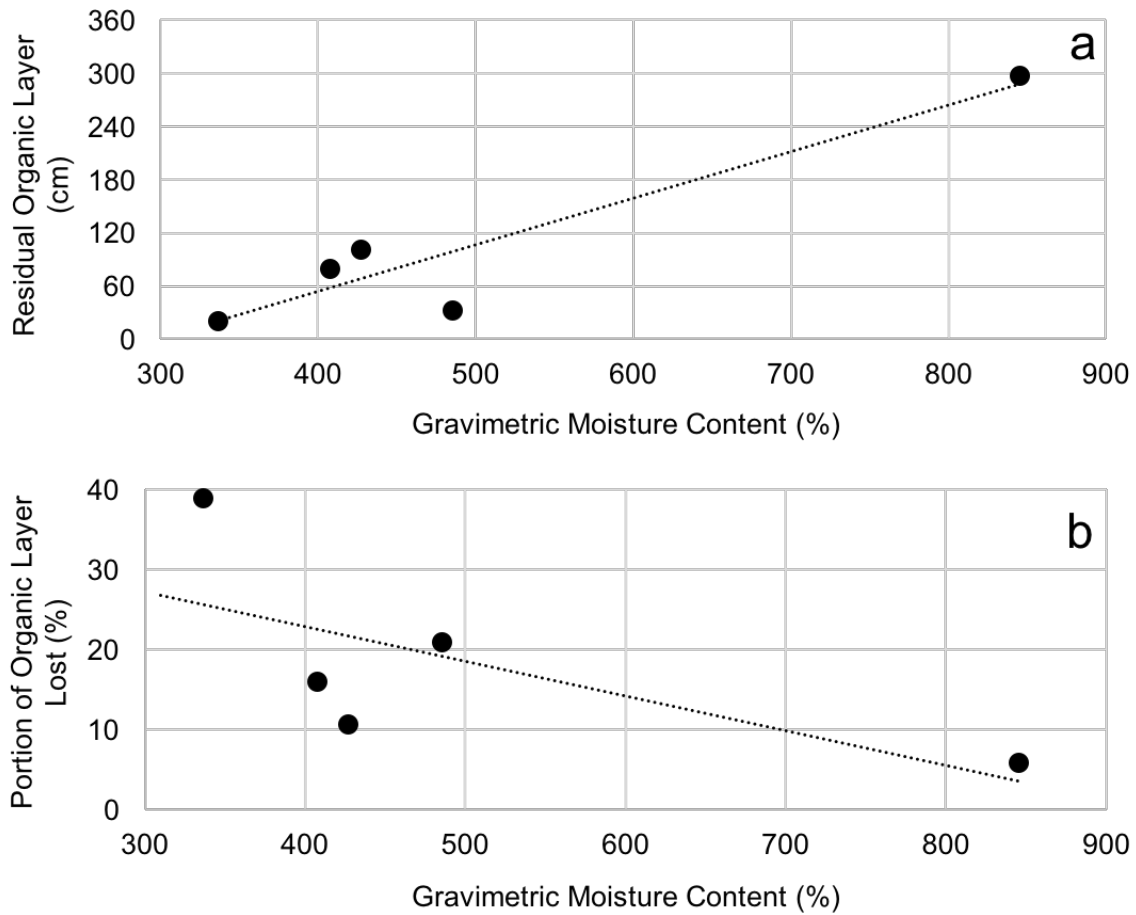


Figure 5-11: Associations between gravimetric moisture content of organic soils in the active layers at burned sites and: (a) residual organic layer thickness, and (b) the portion of the organic layer lost during the fire. Linear relationships were fit to the data, giving an R^2 of (a) 0.88 ($p < 0.05$) and (b) 0.46 ($p > 0.05$). The full range of soil characteristics can be found in Table C3.

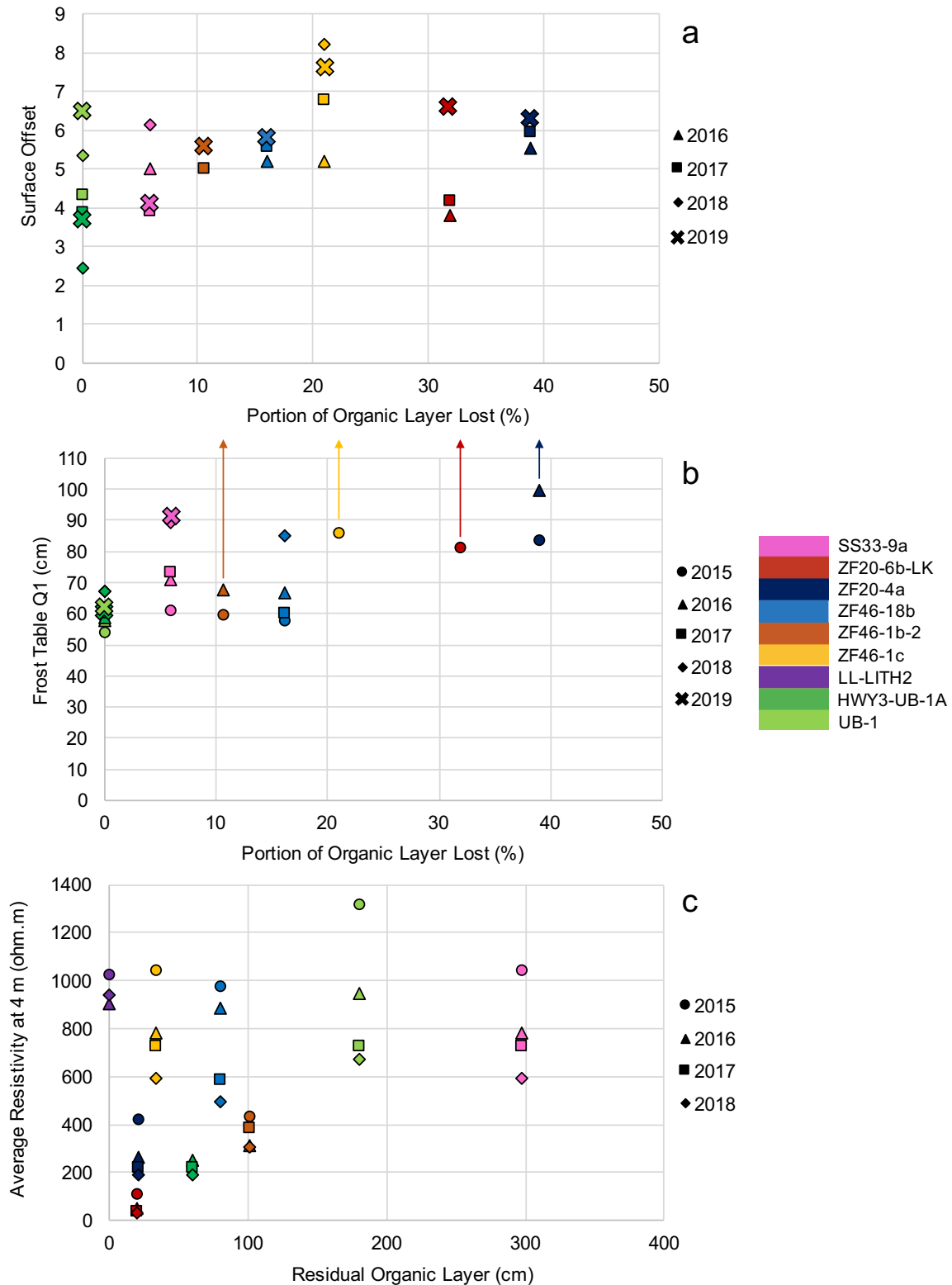


Figure 5-12: Associations between the portion of the organic layer lost during the fire and (a) the surface offset, and (b) the first quartile (Q1) of frost table depths, as well as (c) the relation of average resistivity at 4 m depth and the residual organic layer depth. Note: arrows in (b) indicate that thaw depths in later years exceeded the length of the probe; a single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green.

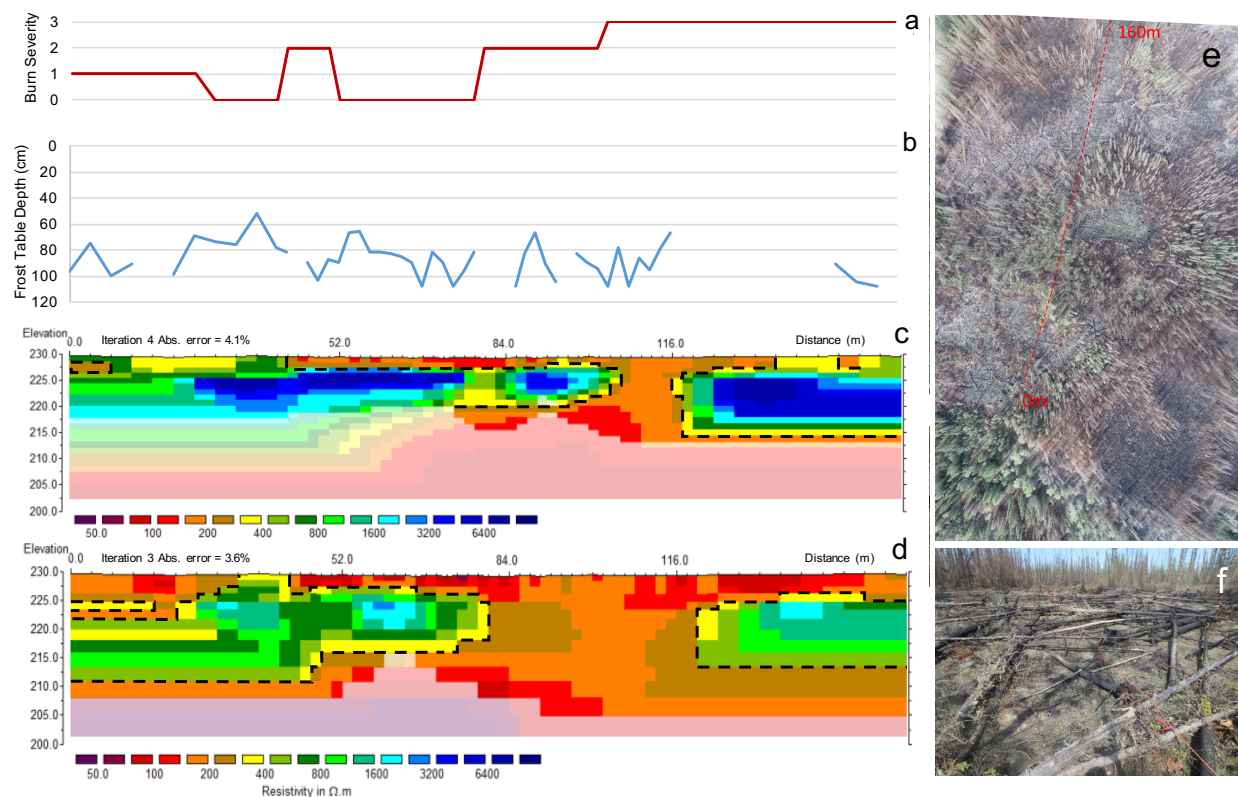


Figure 5-13: Conditions at site ZF20-4a: (a) Descriptive index of burn severity along the transect; (b) frost table depths in September of 2015; ERT surveys shown as model blocks from (c) September 2015 and (d) September 2018; (e) UAV image of site acquired on June 3, 2015; and (f) image of severely burned area from ~120-160 m along the transect acquired on May 14, 2016. Based on frost table measurements, resistivity values $>350 \Omega\text{m}$ are interpreted as frozen ground (dashed line). Note: frost table depths and burn severity were measured at 2-4 m intervals, and blank areas indicate no frost table found within 110 cm of the surface; no frost table was detectable in 2018.

5.5.5 Contribution of ground ice to post-fire permafrost change

The majority of monitoring sites were relatively coarse-grained ($>0.06 \text{ mm}$; see Figures 4-4a and 4-6a). Excess ice was observed in cores taken from five sites, but showed no relationship with post-fire responses of ground temperatures at 1 m depth, frost table depths, and resistivity (Figure C8). Excess ice is known to be present at two additional sites (LL-PEAT and LL-LITH2) but was not quantified.

All eight sites where surface profiles were monitored showed only minor changes in elevation ($< 10 \text{ cm}$) between 2016 and 2018 (Table 5-3). Most showed slight decreases in ground surface level, while two sites apparently heaved by $<5 \text{ cm}$, and one site showed no change at all (Table 5-3). However, all these values fall within the method's estimated error of $\pm 10 \text{ cm}$. The

results indicate, therefore, that essentially no ground subsidence occurred following fire and this includes sites which had excess ice present (Figure C8).

Table 5-3: Difference in ground surface elevation between ZipLevel surveys taken in September, 2016 and September, 2018, calculated by the relative change to anchor points or unburned sections of the profile.

Site	2016-2018 Difference
ZF20-4a	Decrease <5 cm
ZF46-1b-2	Decrease <5 cm
ZF46-1c	Decrease <5 cm
ZF46-18b	Increase <5 cm
LL-PEAT	Decrease <10 cm
LL-LITH2	Decrease <10 cm
HWY3-UB-1A	0 cm
UB-1	Increase <5 cm

Despite the fact that site-wide elevation changes were negligible, sections of a few transects did show more substantial ground subsidence. At ZF46-1b-2, changes in ground surface level of 10-50 cm occurred between 30-60 m along the transect (Figure 5-14a). This area corresponded to a burned peatland, and its subsidence was indicated by the absence of standing water in 2015 and its presence in September 2018 (Figure 5-14b and c). This portion of the transect also experienced a 60% decrease in near-surface resistivity between 2016 and 2018 (Figure C6).

Subsidence was also observed qualitatively in the field at site ZF20-6b-LK where the excess ice content was 18%. The ERT survey in 2015 showed permafrost present from 0-66 m along the transect to a depth of 25 m, and to a depth of about 10 m between 66-100 m (Figure 5-15b). By 2018, frost tables were no longer detectable and changes in resistivities along the transect indicate that the permafrost body between 66-100 m had entirely thawed, while between 0-66 m, thaw depths increased to approximately 5 m, likely including a supra-permafrost talik (Figure 5-15c). This thaw triggered ground subsidence and water inundation around 80 m along the transect between the survey date in 2016 (Figure 5-15d) and 2018 (Figure 5-15e).

Additionally, the borehole which was frozen in 2016 (Figure 5-15f) had thawed out to its maximum depth of 3.2 m by 2019 (Figure 5-15g).

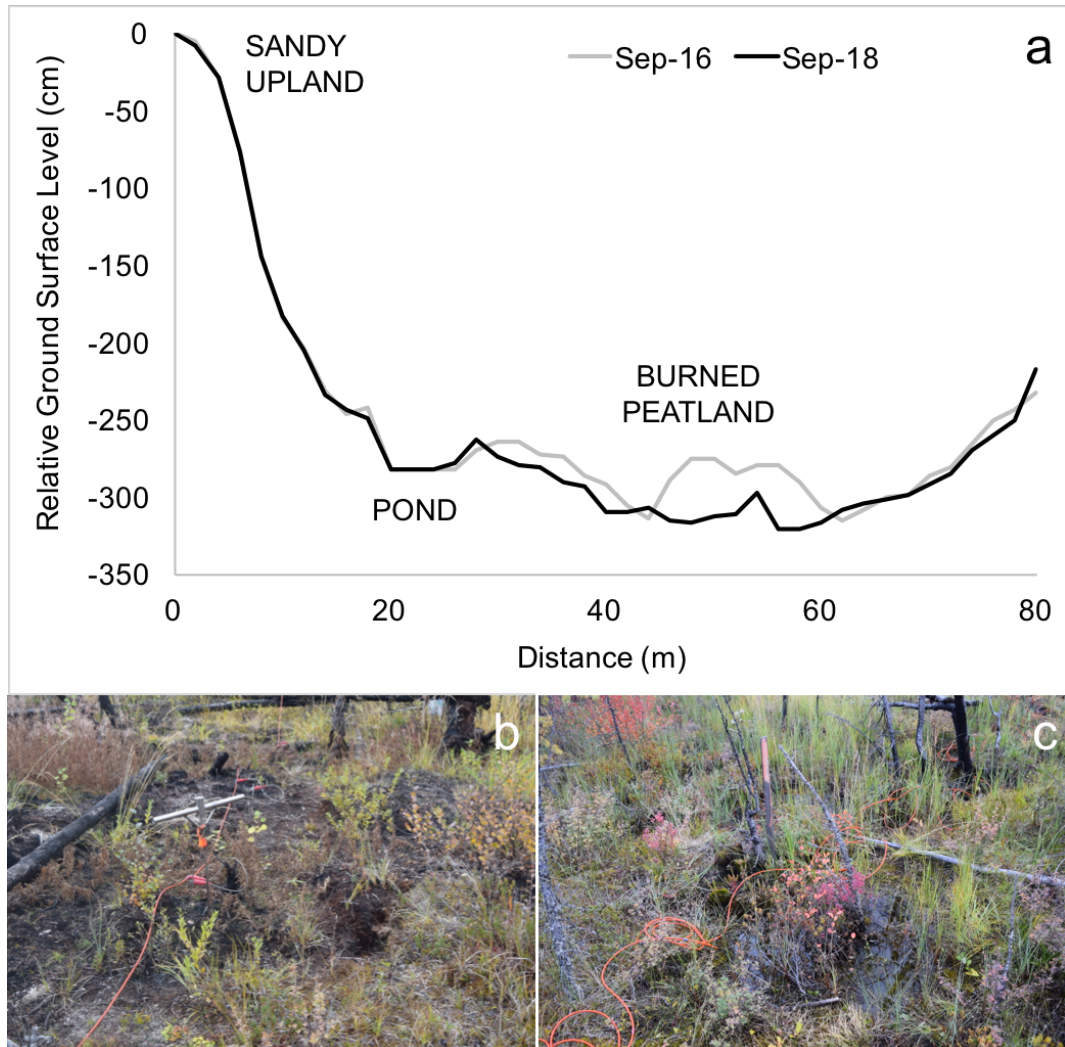


Figure 5-14: Ground subsidence at ZF46-1b2: (a) Difference in ground surface level relative to a stable anchor point (0 m along the transect) between 2016 and 2018 at site ZF46-1b-2; (b) Site at 50 m along the transect in September 2016; and (c) Same area along the transect as in (b) with standing water present in September 2018.

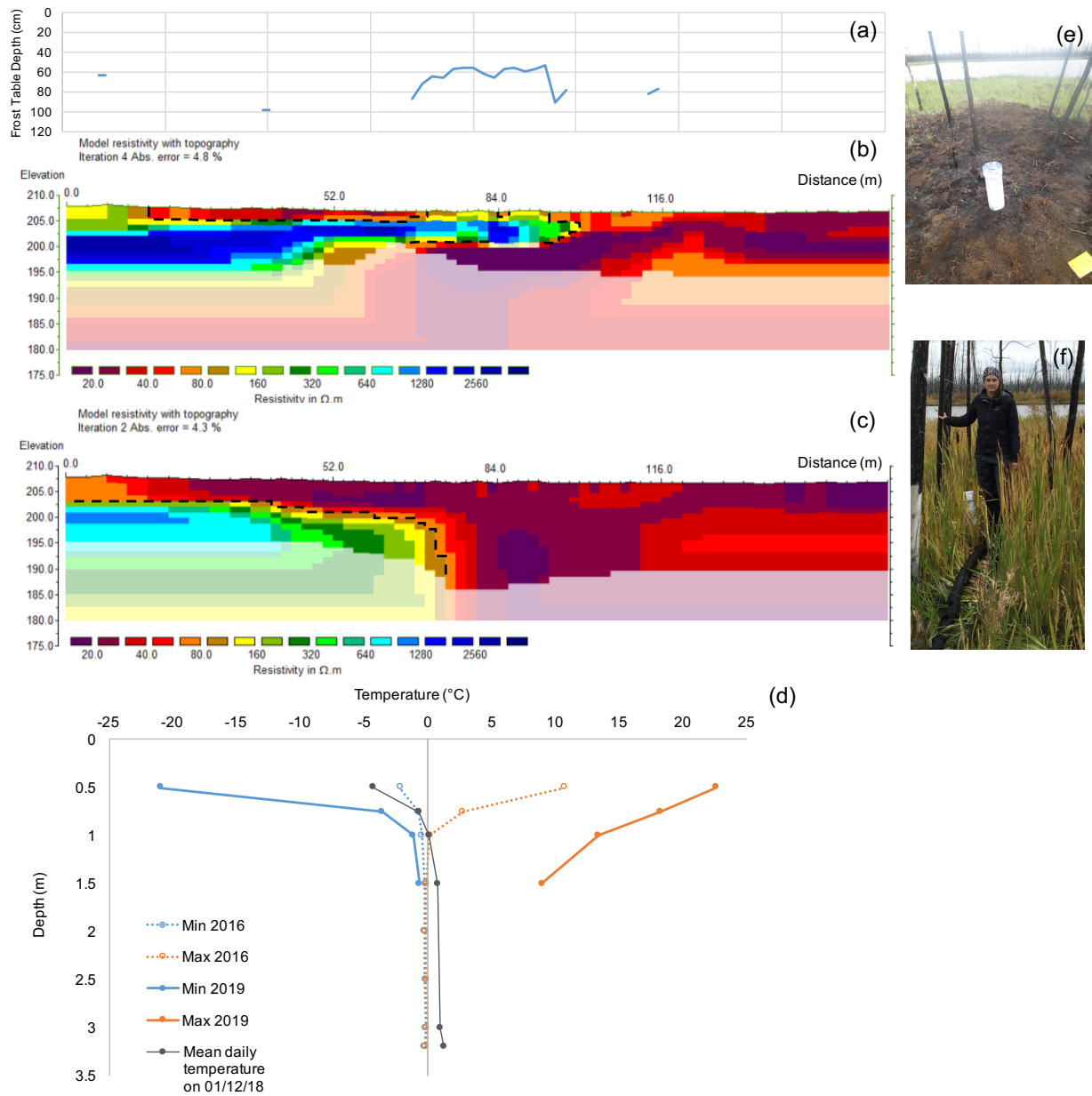


Figure 5-15: Changes in permafrost conditions at site ZF20-6b-LK between 2015 and 2019: (a) Frost table depths in September 2015; (b) ERT survey in September 2015; (c) ERT survey in September 2018; (d) Trumpet curves for borehole temperatures in 2016 and 2019; (e) Photo of the borehole at 80 m along the transect in 2016; and (f) Photo of the same borehole in 2019 after subsidence and inundation. Sensor failure between 2-3.2 m depth occurred for a portion of 2018-2019, but positive temperatures measured down to 3.2 m on December 1, 2018 (black line in (d)) indicate that the borehole had already thawed on that date. Displacement of the inner casing of the borehole may have shifted thermistor depths by ~10 cm between 2015 and 2019. Note: frost table depths were measured at 2-4 m intervals and blank areas indicate no frost table found within 110 cm of the surface; no frost table was detectable in 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded. The threshold between frozen and unfrozen soil for this site, denoted with a dashed line, is 80 Ωm (Holloway and Lewkowicz, 2019).

5.5.6 Post-fire changes in snow and impacts on permafrost

SDD reflect both snowpack depth and the duration of snow cover. At burned sites, average SDD values ranged from 45-74 m.d. Median SDD (averaged for all years) at burned sites was 11% greater than at unburned sites from 2016-2019, but the unburned distribution fell entirely within the IQR of the burned sites (Figure 5-16a). Maximum snow depth (averaged for all years), however, was greater at the burned sites and increased with burn severity, although not significantly (Figure 5-16b). Site ZF46-18b had the greatest average SDD (74 m.d) and maximum snow depth (78 cm).

The nival offset at burned sites was typically higher than at unburned sites (Figure 5-17a). Several of the burned sites showed a progressive increase in nival offset by up to 5.7°C over four years, and up to 3°C at an unburned site. For the unburned site with a complete record, maximum SDD occurred in 2017, whereas for the burned sites three had their maximum SDD in 2018 and two in 2017 (Figure 5-17a). The SDD at site ZF46-18b in 2016 were much greater than any other value observed (Figure 5-17a). This is due to the snow-on date being 22-36 days earlier in that year, as well as a particularly long period (64 days) where the snow remained at the maximum depth (70 cm). Greater frost table depths are associated with greater SDD, and the data suggest that more change may have happened at sites with more snow (Figure 5-17b).

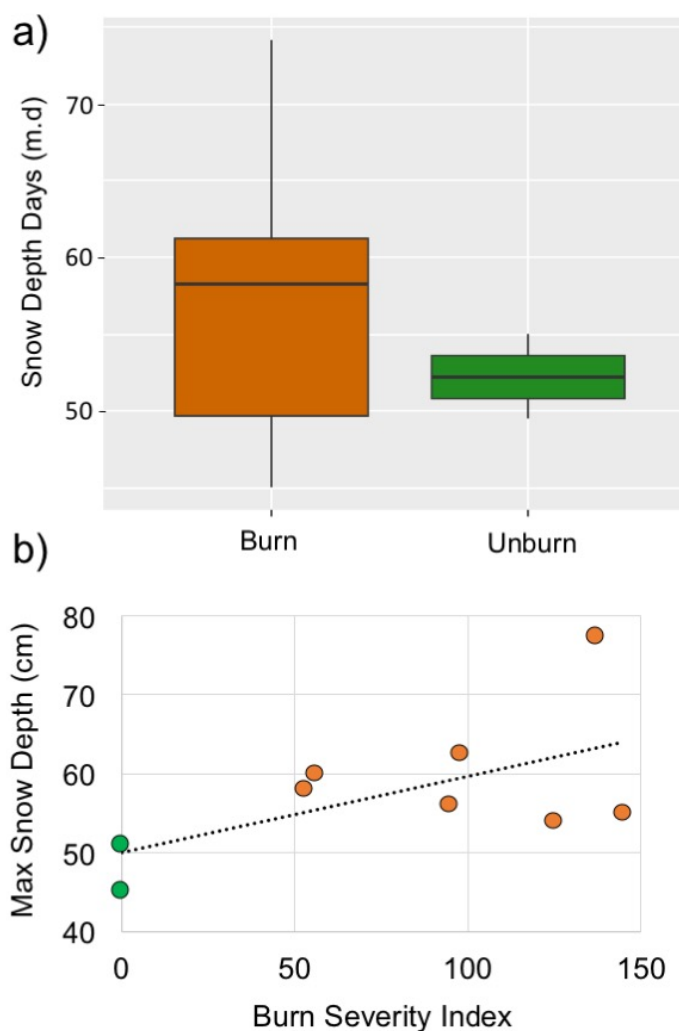


Figure 5-16: Snow conditions at the monitoring sites between 2016-2019: (a) Box-plots of snow-depth days at burned (orange; $n = 7$) and unburned (green; $n = 2$) sites; (b) Average maximum annual snow depth over all years vs. burn severity index ($R^2 = 0.4$, $p > 0.05$). Boxes extend from lower hinge (first quartile) to the upper hinge (third quartile) with the solid line as the median. Whiskers extend to the largest or smallest value no further than $1.5 \times \text{IQR}$ from the hinge (where IQR is the inter-quartile range; Tukey, 1977).

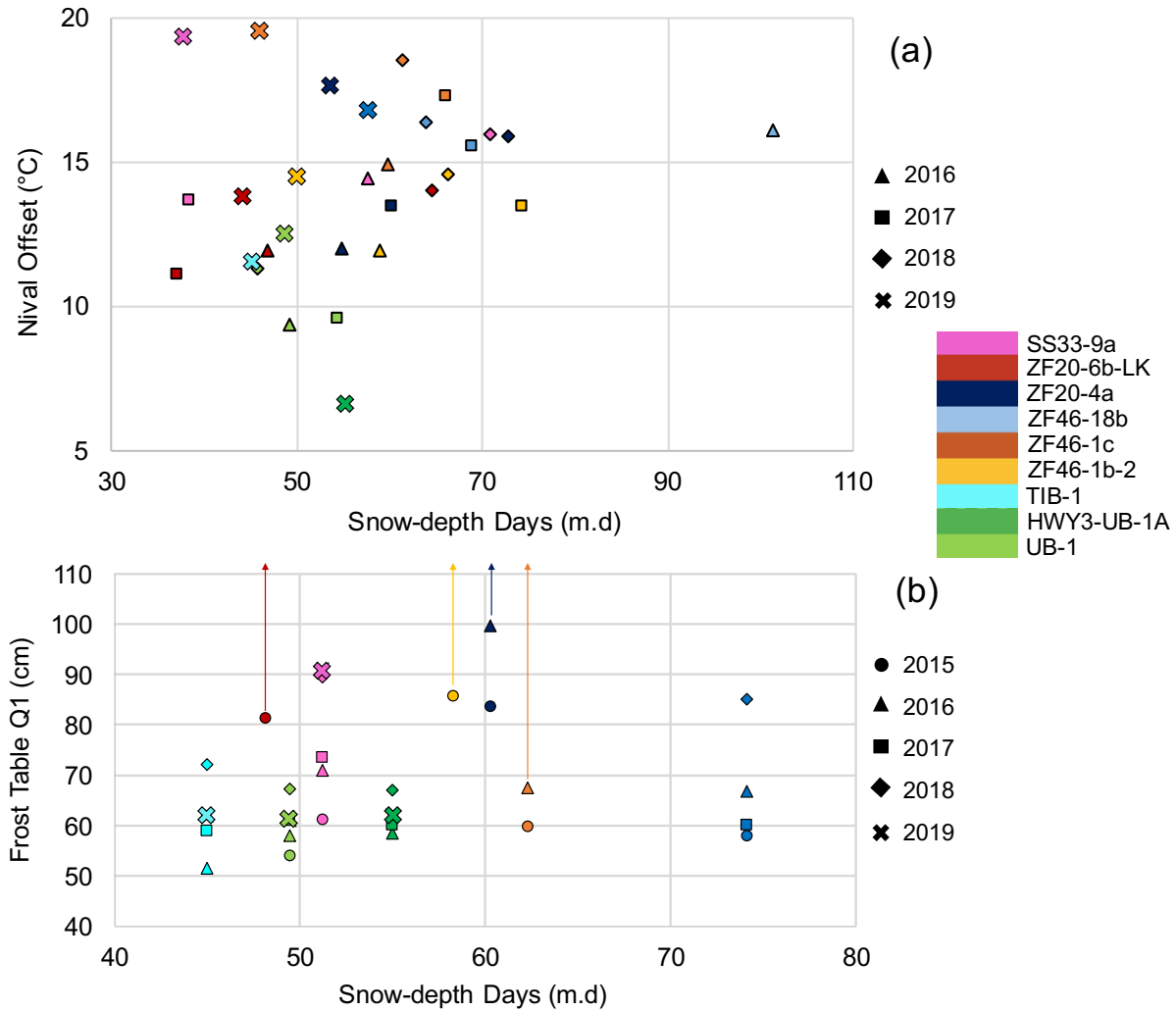


Figure 5-17: Associations between: (a) annual snow-depth days and the nival offset; and (b) the site averaged snow-depth days (2016-2019) and the first quartile (Q1) of frost table depths. Note: arrows indicate that measurements in later years were greater than the length of the probe; a single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green.

5.6 Discussion

This study examined 18 of the 20 factors identified by Holloway et al. (2020) as controls on post-fire permafrost recovery. Information was not collected on post-fire vegetation succession and organic layer recovery, which typically take decades longer than the 5-year period of record. The study observations must be viewed within the context of long-term regional climatic warming of about 2°C over 50 years (Holloway and Lewkowicz, 2020), the fires in 2014 (8 sites) or 2015 (1 site), and a four-year monitoring period from 2015-2019 during which the sites partially revegetated while MAATs progressively decreased by approximately 1.5°C, returning to long-term average values (Figure 5-2). Generalizations are further

complicated by a range of burn severities and antecedent site conditions including soil types, ground ice contents, ground temperatures, soil moisture regimes, and organic layer thicknesses. Finally, instrumentation at most sites was set up in May/June 2015 (Table C1) which was nearly one year after the fires had occurred. In the case of air and ground temperatures, the compiled averages start on September 1, 2015. Changes through time, therefore, are relative to one year after fire and not to pre-fire conditions. Despite these complicating factors, some general patterns were observed (Table 5-4).

Table 5-4: Summary table of site conditions between 2015-2019. Arrows show direction of quantitative changes in mean annual air temperature (MAAT), mean annual ground temperature (MAGT) at 1 m depth, average apparent resistivity at 4 m depth, first quartile (Q1) of frost table (FT) depths, nival offset, and snow-depth days (SDD). The last three columns show average grain size distribution, gravimetric moisture content of organic (O) and mineral (M) soils within the active layer, and the portion of the organic layer (OL) lost in the fire. Sites are organized from south to north, with the unburned sites at the top. Arrows indicate change between years (the first arrow being 2015-2016), with orange representing an increase and blue a decrease, and missing arrows meaning no data was available. For FT depths, a horizontal arrow labelled 110+ is displayed when the Q1 average was greater than the length of the probe. Values for MAGT are given in Table C2. The full grain size distribution was not available for the LL sites, but was observed qualitatively in the field. The full range of soil characteristics can be found in Table C3.

Site	MAAT	MAGT	Resistivity	FT Depth	Nival Offset	SDD (m.d)	Grain size (%)				GMC (%)	% OL Lost
							Gravel	Sand	Silt	Clay		
HWY3-UB-1A (Unburned)	↓ ↓	↓ ↑	↓ ↓	↑ ↑ ↓	↓ ↑	55 in 2018	0	16	68	16	2043 (O)	0
UB-1 (Unburned)	↓ ↓ ↓	↓ ↓ ↑	↓ ↓ ↓	↑ ↑ ↑ ↓	↑ ↑ ↑	↑ ↑ ↑	PEAT				602 (O)	0
SS33-9a (2014)	↓ ↓ ↓	↑ ↑ ↓	↓ ↓ ↓	↑ ↑ ↑ ↑	↓ ↑ ↑	↓ ↑ ↓	0	16	65	19	845 (O)	5
ZF20-6b-LK (2014)	↓ ↓ ↓	↑ ↑	↓ ↓ ↓	↑ 110+	↑ ↑ ↓	↓ ↑ ↓	0	36	57	7	24 (M)	31
ZF20-4a (2014)	↓ ↓ ↓	↑ ↑	↓ ↓ ↓	↑ ↑ 110+	↑ ↑ ↑	↑ ↑ ↓	38	40	19	3	336 (O) 10 (M)	40
ZF46-18b (2014)	↓ ↓ ↓	↑	↓ ↓ ↓	↑ ↑ ↑	↓ ↑ ↑	↓ ↓ ↓	9	67	20	5	407 (O) 7 (M)	16
ZF46-1b-2 (2014)	↓ ↓ ↓	↑ ↑ ↑	↓ ↑ ↓	↑ ↑ 110+	↑ ↑ ↓	↑ ↓ ↓	0	96	4	0	427 (O) 13 (M)	10
ZF46-1c (2014)	↓ ↓ ↓	↓	↓ ↓ ↓	↑ 110+	↑ ↑ ↑	↑ ↓ ↓	8	80	11	1	485 (O) 20 (M)	21
TIB-1 (2015)	↓ ↓	↓ ↑	↑ ↓	↑ ↑ ↓	↓ ↓	45 in 2018	0	20	51	29	310 (O)	
LL-PEAT (2014)	↓ ↑ ↓	↓ ↓	↓ ↓	↑ ↑	↓ ↑ ↑		SILT					
LL-LITH2 (2014)	↓ ↑ ↓	↓ ↑	↓ ↑	↑ ↓	↑ ↓ ↓		SILT					

Based on the literature (Holloway et al., 2020), it was expected that burned sites would exhibit progressive increases in MAGT (Gibson et al., 2018; Jiang et al., 2015; Nossov et al., 2013; O'Donnell et al., 2011b), frost table depths (Fisher et al., 2016; Gibson et al., 2018; Smith et al., 2015; Zhang et al., 2015), and the nival offset, as ground surface temperatures increased in both winter and summer (Jiang et al., 2015; Nossov et al., 2013; Smith et al., 2015). Post-fire permafrost thaw was expected to cause decreases in resistivity, due to a thicker thawed layer and more unfrozen moisture within soil that remained perennially cryotic (Hauck, 2013). These trends were expected to persist over the four-year record, as a return to pre-fire conditions takes decades (Burn, 1998; Gibson et al., 2018; Smith et al., 2015). Permafrost degradation (thaw and warming) due to conductive heat flow from the surface would be greatest at drier sites with thin residual organic layers and coarse textured soil (Fisher et al., 2016; Jafarov et al., 2013; Jiang et al., 2015; Nossov et al., 2013; O'Donnell et al., 2011a; Swanson, 1996; Yi et al., 2010). In contrast, thermokarst development would be restricted to ice-rich sites (Brown et al., 2015; Douglas et al., 2016; Gibson et al., 2018; Olefeldt et al., 2016). All changes would be magnified by increased burn severity (Jafarov et al., 2013; Jiang et al., 2015; Li et al., 2019). Some change was also anticipated at unburned sites as a lag effect of the long-term climate warming, particularly since degradation is ongoing at undisturbed sites in the region (Baltzer et al., 2014; Cannon et al., 2014; Quinton et al., 2011; Wolfe et al., 2015).

MAGT at 1 m depth at burned sites did not progressively increase as had been expected, and the majority of sites exhibited variable temperatures through time (Table 5-4). This is ascribed to the relative strength of the opposing forcings of the lagged impacts of fire and the regional cooling that occurred over the period of record (Figure 5-3). Three of the burned sites did warm, especially ZF20-6b-LK (Figure 5-15d) where, nevertheless, Holloway and Lewkowicz (2019) interpreted frozen ground as persisting to a depth of >20 m. Ground temperatures were warmer at some burned sites compared to unburned ones (Figures 5-5 and 5-17a), but ranges overlapped (Figure 5-3). Interestingly, site LL-PEAT cooled over three years, as did LL-LITH2 (but not between the last two years) (Tables 5-3 and C2). These sites are the northernmost in the study transect (Figure 5-1), and had the lowest MAGT (Table C2). It is possible that the cooling trend could be related to wind scouring of snow due to the relative elevation of the burned lithalsas (Zhang et al., 2015), but this is unlikely due to the presence of a relatively intact birch canopy and rapid understory regrowth. Furthermore, the surface offsets

remain high ($\sim 4\text{--}5^\circ\text{C}$) for these sites (Table C2). Most likely, these trends relate to the limited impact of the fire on the surface energy balance under the deciduous canopy, and less buffering between air and ground temperatures at these cold sites. Therefore, the decline in MAAT over the record resulted in a net negative heat flux and ground cooling. These sites indicate that permafrost vulnerability is affected by antecedent and post-fire regional climate. Overall, significant differences were expected between the thermal regimes at burned and unburned sites, but there was extensive overlap (Figures 5-3 and 5-4).

Resistivity generally behaved as expected, decreasing for most sites between 2015-2018 (Figures 4-8 and Table 5-4), largely in the near-surface. Permafrost at depths greater than about 5 m appeared relatively unaffected but these changes cannot be fully evaluated because deeper layers in the tomograms often showed low sensitivity in the models (Figures 4-9 and 4-15). Percent declines were greater at sites with lower initial resistivity (e.g. Figure 5-8), indicating greater change in unfrozen moisture and that warmer permafrost may be more sensitive to fire. However, at one site with high resistivity (TIB-1), resistivity dropped substantially between 2017 and 2018, likely due to surface water pooling and resulting convective heat transfer and permafrost degradation (Lorant et al., 2018; Neumann et al., 2019). LL-LITH2 experienced a slight increase in resistivity at the end of the record (Table 5-4), and small areas of other transects also showed unexpected increases over time (Figure 5-9). These higher values are probably artefacts of the inversion process but could also represent changes in the moisture content of unfrozen ground. For example, at site ZF20-4a, resistivity increased by $>70\%$ between 76-84 m along the transect at a depth of approximately 10 m (Figure 5-9d). However, this area of the transect had resistivity of $\sim 100\ \Omega\text{m}$ in 2016, which increased to $\sim 200\ \Omega\text{m}$ in 2018. These values are both well below the threshold between frozen and unfrozen soil (Figure 5-13), and may therefore represent a change in the moisture content of the unfrozen soil as the permafrost above thawed.

As expected, frost table depths increased over the four-year record at most sites (Table 5-4), with a few small reversals that are likely due to the decline in MAAT (Figure 5-2). Frost tables at burned sites were generally deeper and changed more over time than at unburned sites (Figure 5-6). As noted above, the only indication of pre-fire conditions comes from the unburned sites. The largest increases in frost table depths happened between 1 and 2 years after fire and the rate of change then slowed (Figure 5-6). These observed post-fire increases are consistent with

results from other studies (Figure 4-18). However, frost table depths observed in this study are on average ~40 cm shallower than those observed elsewhere (Figure 5-18). This may be partly because we measured the minimum active layer thickness at our sites, and frost tables would continue to thicken for several months, making them more comparable with those reported in other literature (Figure B2). Furthermore, we used a 110 cm frost probe and did not include measurements greater than that depth in our sample, contributing to the slight reducing in average frost table depths. Following the broad patterns outlined in Figure 5-18a, frost table depths should continue to increase through to about 2024 (10 years after the fires), followed by gradual recovery to pre-fire levels, although active layers would be thicker due to intervening climate warming (Gibson et al., 2018; Smith et al., 2015; Viereck et al., 2009). As outlined by Holloway et al. (2020), this trajectory would change if thermokarst or talik development occurs at the sites, which has been limited between 2015 and 2019 (Table 5-3). Supra-permafrost and through-going taliks developed at some warmer sites, particularly those that experienced severe burning (e.g. Figures 5-14 and 5-17), and have also been observed elsewhere following fire (Gibson et al., 2018; Viereck et al., 2009). The generally low ice contents meant that visible thermokarst developed at only four sites (e.g. Figures 5-14 and 5-15).

Permafrost degradation, as represented by the measured indicators, was greatest at sites with thin organic layers, coarse textured soil, and lower active layer gravimetric moisture content (Table 5-4). This was apparent at site ZF46-1c, for example, which had a residual organic layer of 33 cm, sandy soil, and average gravimetric moisture content of 20% (e.g. Figures 5-5 and 5-6). Permafrost at this site degraded entirely, with no measurable frost table and average resistivity dropping below 200 Ωm in 2018 (Figures 5-6, 5-8, and C11). However, these major subsurface changes did not lead to detectable surface settlement (Table 5-3), likely due to negligible excess ice associated with the coarse soil texture. In contrast, fine-grained, ice-rich sites underwent ground subsidence following the fire, leading to partial or complete inundation and rapid thaw (e.g., ZF20-6b-LK; Figure 5-15). This type of thermokarst following fire has been observed elsewhere in the region, although mostly in peatlands (Gibson et al., 2018). However, measurable subsidence did not occur at all burned sites with excess ice. LL-PEAT and LL-LITH2 were only minimally impacted (Table 5-3 and Figure C8), likely because these sites remained cold (Figure 5-5 and Table C2). Regardless, soil texture and ground ice content were strong determinants for post-fire permafrost trajectory (Holloway et al., 2020; Nossor et al., 2013). More work is needed to quantify impacts of fire on low ice-content permafrost sites with coarse-grained soils, since our results show these landscapes can be sensitive to

fire, and a large portion of permafrost in the discontinuous zone is of this type (Olefeldt et al., 2016; O'Neill et al., 2019; 2020).

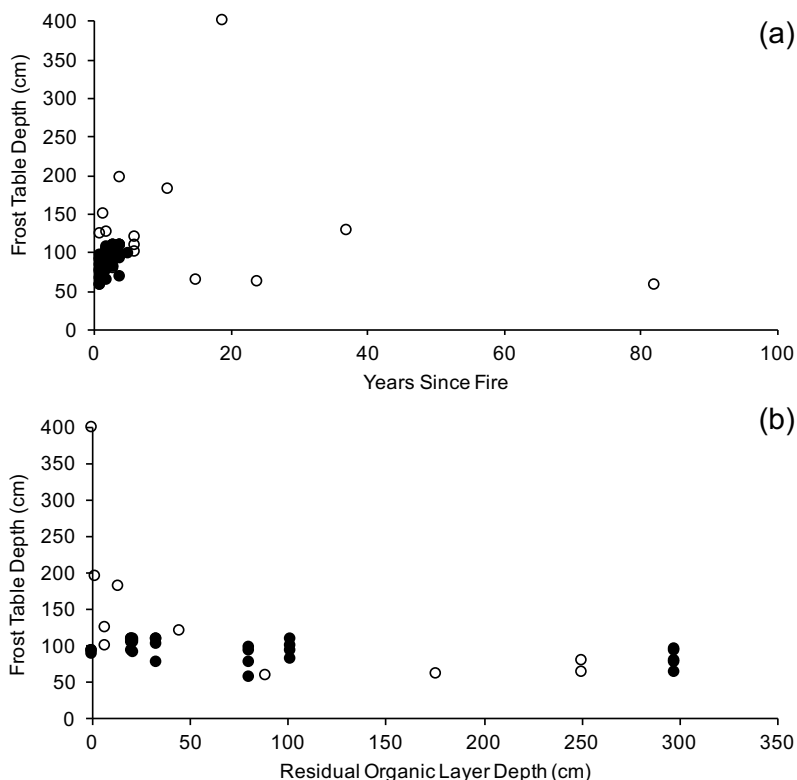


Figure 5-18: Average post-fire frost table depths compared to (a) years since fire and (b) the residual organic layer depth measured in this study (black) and in recent literature (white) at sites in the boreal forest (Fisher et al., 2016; Gibson et al., 2018; Jafarov et al., 2013; Jiang et al., 2015; Li et al., 2019; Nossor et al., 2013; Zhang et al., 2015). Figure modified from Holloway et al. (2020). Note: for measurements from this study, points where the frost table was greater than the 110 cm probe have been included in the sample.

Increased burn severity did result in more permafrost change over time (Table 5-4). Sites which experienced the most change over time, particularly increased frost table depths, had a greater portion of the organic layer lost during the fire (Figure 5-12b). However, compared to the literature, frost tables were much shallower where residual organic layers were thin (Figure 5-18b). As mentioned above, this is likely due to antecedent ground temperatures, as sites at the northern end of the study transect are colder (MAGT of $\sim -2^{\circ}\text{C}$) than most that have been monitored following fire in the North American boreal forest (MAGT of $> -1^{\circ}\text{C}$; e.g. Nossor et al., 2013; Smith et al., 2015). Residual organic layers were deeper at sites with higher active layer moisture content, and less of the organic layer was consumed during the fire at wet sites

(Figure 5-11), consistent with literature that shows burning is more intense at drier sites (Dillon et al., 2011; Turetsky et al., 2011; Walker et al., 2018a).

To date, there have been no direct observations of the impacts that post-fire changes in snow have on permafrost (Holloway et al., 2020). Lower interception due to loss of the canopy following fire and increased snow depth is suspected to be the cause of higher ground temperatures at sites in the NWT (Smith et al., 2015). In non-permafrost landscapes, observations suggest that the snowpack in open, recovering forests is also affected by greater redistribution and compaction by wind (Burles and Boon, 2011). In addition, ablation rates are accelerated and snow disappears days to weeks earlier due to the decreased albedo of the snow associated with charred debris, as well as more solar radiation reaching the surface (Burles and Boon, 2011; Gleason et al., 2013). Ultimately, SDD and maximum snow depths were expected to be highest at burned sites in 2015 and to decrease over the record as revegetation occurred, while conforming to the pattern of inter-annual fluctuations in snow measured at the unburned sites. However, SDD at burned sites did not behave as anticipated (Table 5-4). Decreases in SDD over time were observed at some of the burned sites, but others showed fluctuations that were inconsistent with the pattern at the unburned sites (Table 5-4). However, median SDD (Figure 5-16a), maximum snow depth (Figure 5-16b), and the nival offset (Figure 5-17a) were all greater at burned compared to unburned sites. Note, however, that due to the limited number of unburned sites these comparisons are tentative. Snow depth increased with burn severity, although the relationship was not significant (Figure 5-17b). At sites where the nival offset increased progressively through the period of record (ZF20-4a, ZF46-1c), SDD did not increase simultaneously (Figure 5-17a and Table 5-4), which suggests surface temperatures were also being affected by latent heat released during ground thaw over that time (Figure 5-4, 5-6, and Table C2) (Riseborough and Smith, 1998).

The sites in this study span a large swath of permafrost in the western Canadian boreal forest (Figure 5-1). Our initial hypothesis was that permafrost at relatively warm sites farther south ($\text{MAAT} > -2^{\circ}\text{C}$) would experience greater post-fire change than colder sites at the northern end of the transect ($\text{MAAT} < -2^{\circ}\text{C}$), due to the loss of ecosystem protection. However, the response of frozen ground following fire did not follow a latitudinal gradient (Figure 5-19). Changes in frost table depths in the years following fire occurred at sites across the range of latitudes, with large increases at sites with annual air temperatures for the period averaging as

low as -3.6°C . Further, the only site that experienced complete post-fire permafrost degradation (ZF46-1c) is located in the middle of the transect at the same average MAAT (Figure 5-1 and Table C1). In contrast, the most southerly site (SS33-9a) showed only moderate increases in frost table depths, likely due to its thick residual organic layer (~ 3 m). These observations show that permafrost at the monitoring sites where MAAT is warmer than -2°C should not necessarily be classified as ecosystem-protected since it persisted, at least in the short-term, following fire (Jorgenson et al., 2010). Further monitoring or thermal modelling would be needed to evaluate its long-term behaviour. Measurements south of this study transect indicate that there is little difference between burned and unburned frost table depths 6-47 years post-fire (Holloway and Lewkowicz, 2020). It is clear that the trajectory followed by warm permafrost after a fire depends more on site-specific characteristics, such as soil type, ground ice conditions, and residual organic layer thickness, than on latitude or regional permafrost zone. However, colder sites at the northern end of the transect appear more resilient to fire due to lower antecedent ground temperatures, even where thin residual organic layers are present (e.g. LL-LITH2; Figures 5-5, 5-8, and 5-9).

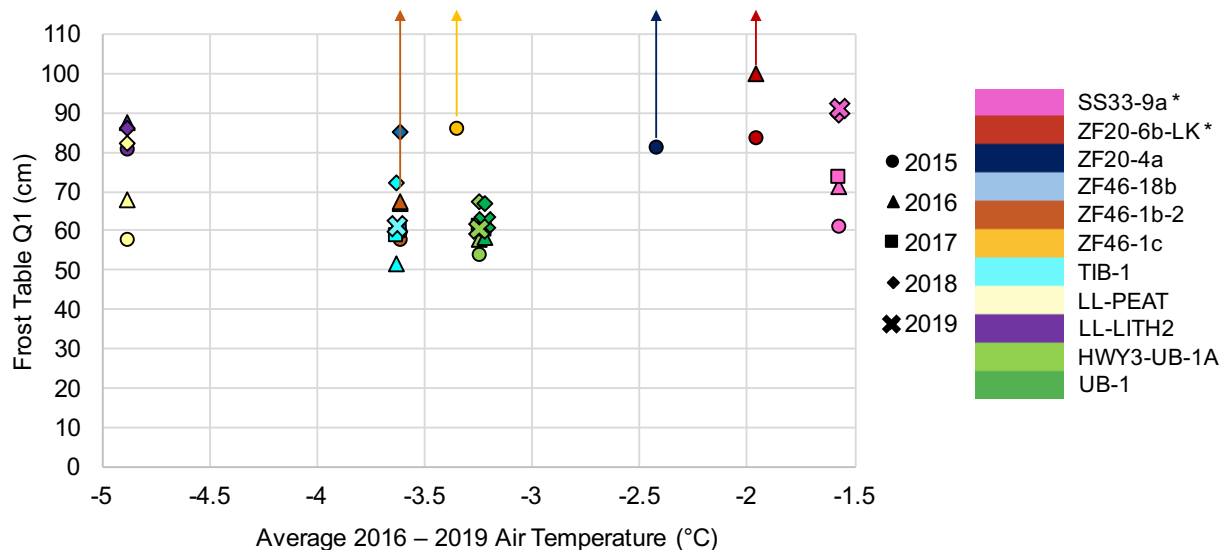


Figure 5-19: Changes in the first quartile (Q1) of frost table depths compared to average air temperature from 2015-2019. Note: arrows indicate that measurements in later years were greater than the length of the probe; a single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green. Sites where permafrost could be ecosystem-protected (MAAT $> -2^{\circ}\text{C}$) are designated with an asterisk (*).

5.7 Conclusion

Permafrost was degrading at the start of monitoring in the southern NWT as a result of ongoing regional climate warming, but was exacerbated at those forested sites which burned in 2014 and 2015, as shown by progressive reductions in electrical resistivity and increases in frost table depths and ground temperatures. This occurred in spite of annual air temperatures that declined in the region by $\sim 2^{\circ}\text{C}$ during the actual monitoring period (2015-2019). Permafrost change was predominantly in the near-surface in the five years post-fire. Degradation was more pronounced at sites with low gravimetric moisture contents, coarse soils, and thin organic layers (<40 cm), except at the small number of sites where thaw settlement and surface inundation occurred. Supra-permafrost and through-going taliks developed at some warmer sites, particularly those that experienced severe burning. SDD, maximum snow depth, and the nival offset were all greater at burned compared to unburned sites.

Permafrost at colder sites at the northern end of the transect, as well as permafrost under thick organic layers at the southern end, appeared resilient to fire, whereas sites with thin organic layers and coarse soils on the taiga plains were the most sensitive. This demonstrates that warm permafrost in southern locations can remain resilient to fire (at least in the short-term) and emphasizes the importance of monitoring sites over a wide range of environmental conditions. In particular, low ice-content permafrost sites need further evaluation as a large portion of the boreal zone is covered by these soils, and their reaction to fire has implications for hydrology, biogeochemistry and carbon cycling. This is especially important as the frequency, severity, and extent of wildfire is predicted to increase with climate warming. It is essential to incorporate the impact of wildfire on permafrost landscapes into regional modelling of long-term change.

Previous studies of the impact of fire on permafrost led us to expect clear, straightforward, consistent results. We hypothesized substantial differences between burned and unburned sites, and that fire would lead to permafrost thaw and degradation, especially at southern sites, where the burn was severe, and the ground was ice-rich. However, the signal proved mixed at nine burned sites and we conclude that site heterogeneity can trump the disturbance due to fire. While fire has a clear impact on the vegetation of the boreal forest, its effects on permafrost are often harder to discern, and some perennially frozen ground remains resilient, even in the southernmost reaches of permafrost in western Canada.

5.8 References

- Baltzer, J. L., Veness, T., Chasmer, L. E., Sniderhan, A. E., and Quinton, W. L. (2014). Forests on thawing permafrost: fragmentation, edge effects, and net forest loss. *Global Change Biology*, 20(3), 824-834. <https://doi.org/10.1111/gcb.12349>
- Beilman, D. W., and Robinson, S. D. (2003). Recent permafrost melt along a climatic gradient in western Canada: large-scale peat mapping. In M. Phillips, S. M. Springman, L. U. Arenson, A. A. Balkema (Eds.), *Proceedings of the 8th International Conference on Permafrost* (pp. 61-65). Zurich, Switzerland.
- Boby, L. A., Schuur, E. A., Mack, M. C., Verbyla, D., and Johnstone, J. F. (2010). Quantifying fire severity, carbon, and nitrogen emissions in Alaska's boreal forest. *Ecological Applications*, 20, 1633-1647. <https://doi.org/10.1890/08-2295.1>
- Brown, J., Hinkel, K. M., and Nelson, F. E. (2000) The Circumpolar Active Layer Monitoring (CALM) program: Research designs and initial results. *Polar Geography*, 24(3), 165–258.
- Brown, D. R. N., Jorgenson, M. T., Douglas, T. A., Vladimir, R. E., Kielland, K., Hiemstra, C., Euskirchen, E. S., and Ruess, R. W. (2015). Interactive effects of wildfire and climate on permafrost degradation in Alaskan lowland forests. *Journal of Geophysical Research - Biogeosciences*, 120(8), 1619-1637. <https://doi.org/10.1002/2015JG003033>
- Burles, K., and Boon, S. (2011). Snowmelt energy balance in a burned forest plot, Crowsnest Pass, Alberta, Canada. *Hydrological Processes*, 25(19), 3012-3029. <https://doi.org/10.1002/hyp.8067>
- Burn, C. R. (1998). The response (1958-1997) of permafrost and near-surface ground temperatures to forest fire, Takhini River Valley, southern Yukon Territory. *Canadian Journal of Earth Science*, 35(2), 184-199. <https://doi.org/10.1139/e97-105>
- Canadian Interagency Forest Fire Center. (2014). *Situation Report – Sep 22, 2014*. <http://www.cifffc.ca/firewire/current.php?lang=en&date=20140922>
- Connon, R. F., Quinton, W. L., Craig, J. R., and Hayashi, M. (2014). Changing hydrologic connectivity due to permafrost thaw in the lower Liard River valley, NWT, Canada. *Hydrological Processes*, 28(14), 4163-4178. <https://doi.org/10.1002/hyp.10206>
- de Groot, W. J., Flannigan, M. D., and Cantin, A. S. (2013). Climate change impacts on future boreal fire regimes. *Forest Ecology and Management*, 294, 35-44. <https://doi.org/10.1016/j.foreco.2012.09.027>
- Dillon, G. K., Holden, Z. A., Morgan, P., Crimmins, M. A., Heyerdahl, E. K., and Luce, C. H. (2011). Both topography and climate affected forest and woodland burn severity in two regions of the western US, 1984 to 2006. *Ecosphere*, 2(12), 1-33. <https://doi.org/10.1890/ES11-00271.1>

- Douglas, T. A., Jorgenson, M. T., Brown, D. R., Campbell, S. W., Hiemstra, C. A., Saari, S. P., Bjella, K., and Liljedahl, A. K. (2016). Degrading permafrost mapped with electrical resistivity tomography, airborne imagery and LiDAR, and seasonal thaw measurements. *Geophysics*, 81(1), 1JF-Z7. <https://doi.org/10.1190/geo2015-0149.1>
- Ecological Stratification Working Group. (1996). *A National Ecological Framework for Canada*. Centre for Land and Biological Resources Research, Research Branch, Agriculture and Agri-Food Canada.
- Environment Canada. (2019). *Historical climate data*. http://climate.weather.gc.ca/historical_data/search_historic_data_e.html.
- Fisher, J. P., Estop-Aragónés, C., Thierry, A., Charman, D. J., Wolfe, S. A., Hartley, I. P., Murton, J. B., Williams, M., and Phoenix, G. K. (2016). The influence of vegetation and soil characteristics on active-layer thickness of permafrost soils in boreal forest. *Global Change Biology*, 22, 3127-3140. <https://doi.org/10.1111/gcb.13248>
- French, H. M. (2018). *The Periglacial Environment* (4th ed.). John Wiley & Sons, Ltd.
- Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., and Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nature Communications*, 9, 3041. <https://doi.org/10.1038/s41467-018-05457-1>
- Gleason, K. E., Nolin, A. W., and Roth, T. R. (2013). Charred forests increase snowmelt: effects of burned woody debris and incoming solar radiation on snow ablation. *Geophysical Research Letters*, 40(17), 4654-4661. <https://doi.org/10.1002/grl.50896>
- Hanes, C. C., Wang, X., Jain, P., Parisien, M., Little, J. M., and Flannigan, M. D. (2019). Fire-regime changes in Canada over the last half century. *Canadian Journal of Forest Research*, 49(3), 256-269. <https://doi.org/10.1139/cjfr-2018-0293>
- Hauck, C. (2013). New concepts in geophysical surveying and data interpretation for permafrost terrain. *Permafrost and Periglacial Processes*, 24(2), 131-137. <https://doi.org/10.1002/ppp.1774>
- Heginbottom, J. A., Dubreuil, M. A., and Harker, P. A. (1995). Canada – Permafrost. *National Atlas of Canada* (5th ed.). Natural Resources Canada.
- Heiri, O., Lotter, A.F., and Lemcke, G. (2001). Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. *Journal of Paleolimnology*, 25(1), 101-110. <https://doi.org/10.1023/A:1008119611481>
- Holloway, J., and Lewkowicz, A. G. (2019). Field and laboratory investigation of electrical resistivity-temperature relationships, southern Northwest Territories. In J.-P. Bilodeau, D.F. Nadeau, D. Fortier, and D. Conciatori (eds.), *Cold Regions Engineering 2019: Proceedings of the 18th International Conference on Cold Regions Engineering and 8th*

- Canadian Permafrost Conference* (pp. 64-72). American Society of Civil Engineers, Reston, Virginia. <https://doi.org/10.1061/9780784482599>
- Holloway, J. E., and Lewkowicz, A. G. (2020). Persistence of permafrost following 55 years of climate warming and forest fire in the discontinuous zone, Canada. *Permafrost and Periglacial Processes*, 31(1), 85-96. <https://doi.org/10.1002/ppp.2017>
- Holloway, J. E., Lewkowicz, A. G., Douglas, T. A., Li, X., Turetsky, M. R., Baltzer, J. L., Jin, H. (2020). The impact of wildfire on permafrost landscapes: a review of recent advances and future prospects. *Permafrost and Periglacial Processes*, 31(3), 371-382. <https://doi.org/10.1002/ppp.2048>
- Jafarov, E. E., Romanovsky, V. E., Genet, H., McGuire, A. D., and Marchenko, S. S. (2013). The effects of fire on the thermal stability of permafrost in lowland and upland black spruce forests of interior Alaska in a changing climate. *Environmental Research Letters*, 8(3), 035030. <http://dx.doi.org/10.1088/1748-9326/8/3/035030>
- Jiang, Y., Rocha, A. V., O'Donnell, J. A., Drysdale, J., Rastetter, E. B., Shaver, G. R., and Zhuang, Q. (2015). Contrasting soil thermal responses to fire in Alaskan tundra and boreal forest. *Journal of Geophysical Research: Earth Surface*, 120(2), 363-378. <http://dx.doi.org/10.1002/2014JF003180>
- Jones, M. C., Grosse, G., Jones, B. M., and Anthony, K. W. (2012). Peat accumulation in drained thermokarst lake basins in continuous, ice-rich permafrost, northern Seward Peninsula, Alaska. *Journal of Geophysical Research: Biogeosciences*, 117(G2). <https://doi.org/10.1029/2011JG001766>
- Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., O'Donnell, J., Schuur, E. A. G., Kanevskiy, M., and Marchenko, S. (2010). Resilience and vulnerability of permafrost to climate change. *Canadian Journal of Forest Research*, 40(7), 1219-1236. <https://doi.org/10.1139/X10-060>
- Kasischke, E.S., and Johnstone, J.F. (2005). Variation in postfire organic layer thickness in a black spruce forest complex in interior Alaska and its effects on soil temperature and moisture. *Canadian Journal of Forest Research*, 35, 2164-2177. <https://doi.org/10.1139/X05-159>
- Kasischke, E. S., Verbyla, D. L., Rupp, T. S., McGuire, A. D., Murphy, K. A., Jandt, R., Barnes, J. L., Hoy, E. E., Duffy, P. A., Calef, M., and Turetsky, M. R. (2010). Alaska's changing fire regime - implications for the vulnerability of its boreal forests. *Canadian Journal of Forest Research*, 40(7), 1313-1324. <https://doi.org/10.1139/X10-098>
- Knorre, A. A., Kirdyanov, A. V., Prokushkin, A. S., Krusic, P. J., Buntgen, U. (2019). Tree ring-based reconstruction of the long-term influence of wildfires on permafrost active layer dynamics in Central Siberia. *Science of the Total Environment*, 652, 314-319. <https://doi.org/10.1016/j.scitotenv.2018.10.124>

- Kokelj, S. V., and Burn, C. R. (2003). Ground ice and soluble cations in near-surface permafrost, Inuvik, Northwest Territories, Canada. *Permafrost and Periglacial Processes*, 14(3), 275-289.
- Lewkowicz, A. G. (2008). Evaluation of miniature temperature-loggers to monitor snowpack evolution at mountain permafrost sites, northwestern Canada. *Permafrost and Periglacial Processes*, 19(3), 323-331.
- Lewkowicz, A. G., Etzelmüller, B., and Smith, S. L. (2011). Characteristics of discontinuous permafrost based on ground temperature measurements and electrical resistivity tomography, southern Yukon, Canada. *Permafrost and Periglacial Processes*, 22(4), 320-342. <https://doi.org/10.1002/ppp.703>
- Li, X., Jin, H., He, R., Huang, Y., Wang, H., Luo, D., Jin, X., Lü, L., Wang, L., Li, W., Wei, C., Chang, X., Yang, S., and Yu, S. (2019). Effects of forest fires on the permafrost environment in the northern Da Xing'anling (Hinggan) mountains, Northeast China. *Permafrost and Periglacial Processes*, 30(3), 163-177. <https://doi.org/10.1002/ppp.2001>
- Loke, M. H. (2001). *Electrical imaging surveys for environmental and engineering studies, a practical guide to 2-D and 3-D surveys*. RES2DINV Manual, IRIS Instruments.
- Loke, M. H., Acworth, I., and Dahlin, T. (2003). A comparison of smooth and blocky inversion methods in 2D electrical imaging surveys. *Exploration Geophysics*, 34(3), 182–187. <https://doi.org/10.1071/EG03182>
- Loranty, M. M., Abbott, B. W., Blok, D., Douglas, T. A., Epstein, H. E., Forbes, B. C., Jones, B. M., Kholodov, A. L., Kropp, H., Malhotra, A., Mamet, S. D., Myers-Smith, I. H., Natali, S. M., O'Donnell, J. A., Phoenix, G. K., Rocha, A. V., Sonnentag, O., Tape, K. D., and Walker, A. D. (2018). Reviews and syntheses: Changing ecosystem influences on soil thermal regimes in northern high-latitude permafrost regions. *Biogeosciences*, 15(17), 5287-5313. <https://doi.org/10.5194/bg-15-5287-2018>
- Mackay, J. R. (1977). Probing for the bottom of the active layer. Geological Survey of Canada, Paper No. 77-1A, 327-328.
- Miller, C. R., Routh, P. S., Brosten, T. R., McNamara, J. P. (2008). Application of time-lapse ERT imaging to watershed characterization. *Geophysics*, 73(3), G7–G17. <https://doi.org/10.1190/1.2907156>
- Minsley, B. J., Pastick, N. J., Wylie, B. K., Brown, D. R. N., and Kass, M. A. (2016). Evidence for nonuniform permafrost degradation after fire in boreal landscapes. *Journal of Geophysical Research: Earth Surface*, 121(2), 320-335. <https://doi.org/10.1002/2015JF003781>
- Morse, P. D., Burn, C. R., and Kokelj, S. V. (2012). Influence of snow on near-surface ground temperatures in upland and alluvial environments of the outer Mackenzie Delta,

- Northwest Territories. *Canadian Journal of Earth Sciences*, 49(8), 895–913. <https://doi.org/10.1139/e2012-012>
- Natural Resources Canada. (2020). *Precise Point Positioning*. Canadian Geodetic Survey. <https://webapp.geod.nrcan.gc.ca/geod/tools-outils/ppp.php?locale=en>
- Neumann, R. B., Moorberg, C. J., Lundquist, J. D., Turner, J. C., Waldrop, M. P., McFarland, J. W., Euskirchen, E. S., Edgar, C. W., Turetsky, M. R. (2019). Warming effects of spring rainfall increase methane emissions from thawing permafrost. *Geophysical Research Letters*, 46(3), 1393-1401. <https://doi.org/10.1029/2018GL081274>
- Northern Climate Exchange. (2013). *Burwash Landing and Destruction Bay landscape hazards: geological mapping for climate change adaptation planning*. Yukon Research Centre, Yukon College. https://www.yukonu.ca/sites/default/files/inline-files/Hazards_BurwashDBay_finalApr29_web.pdf
- Nossov, D. R., Jorgenson, M. T., Kielland, K., and Kanevskiy, M. Z. (2013). Edaphic and microclimatic controls over permafrost response to fire in interior Alaska. *Environmental Research Letters*, 8(3), 035013. <https://doi.org/10.1088/1748-9326/8/3/035013>
- O'Donnell, J. A., Harden, J. W., McGuire, A. D., Kanevskiys, M. Z., Jorgenson, M. T., and Xu, X. (2011a). The effect of fire and permafrost interactions on soil carbon accumulation in an upland black spruce ecosystem of interior Alaska: implications for post-thaw carbon loss. *Global Change Biology*, 17(3), 1461-1474. <https://doi.org/10.1111/j.1365-2486.2010.02358.x>
- O'Donnell, J. A., Harden, J. W., McGuire, A. D., and Romanovsky, V. E. (2011b). Exploring the sensitivity of soil carbon dynamics to climate change, fire disturbance and permafrost thaw in a black spruce ecosystem. *Biogeosciences*, 8, 1367-1382. <https://doi.org/10.5194/bg-8-1367-2011>
- O'Neill, H. B., Wolfe, S. A., and Duchesne, C. (2019). New ground ice maps for Canada using a paleogeographic modelling approach. *Cryosphere*, 13, 753–773. <https://doi.org/10.5194/tc-13-753-2019>
- O'Neill, H. B., Wolfe, S. A., and Duchesne, C. (2020). *Ground ice map of Canada v.1*. Geological Survey of Canada, Open File 8713. In press.
- Oldenborger, G. A., Knoll, M. D., Routh, P. S., LaBrecque, D. J. (2007). Time-lapse ERT monitoring of an injection/withdrawal experiment in a shallow unconfined aquifer. *Geophysics*, 72(4), 177–187. <https://doi.org/10.1190/1.2734365>
- Olefeldt, D., Goswami, S., Grosse, G., Hayes, D., Hugelius, G., Kuhry, P., McGuire, A. D., Romanovsky, V. E., Sannel, A. B. K., Schuur, E. A. G., Turetsky, M. R. (2016). Circumpolar distribution and carbon storage of thermokarst landscapes. *Nature Communications*, 7, 13043. <https://doi.org/10.1038/ncomms13043>

- Quinton, W. L., Hayashi, M., and Chasmer, L. E. (2011). Permafrost-thaw-induced landcover change in the Canadian subarctic: Implications for water resources. *Hydrological Processes*, 25(1), 152–158. <https://doi.org/10.1002/hyp.7894>
- Riseborough, D. W., and Burgess, M. M. (1996). Measurement interval and the accurate assessment of ground temperature trends. *Permafrost and Periglacial Processes*, 7(4), 321-336. [https://doi.org/10.1002/\(SICI\)1099-1530\(199610\)7:4%3C321::AID-PPP233%3E3.0.CO;2-J](https://doi.org/10.1002/(SICI)1099-1530(199610)7:4%3C321::AID-PPP233%3E3.0.CO;2-J)
- Riseborough, D. W., and Smith, M. W. (1998). Exploring the limits of permafrost. In A. G. Lewkowicz, M. Allard (eds.), *Proceedings of Seventh International Conference on Permafrost* (pp. 935-941). Centre d'études Nordiques, Collection Nordicana, 57.
- Smith, M. W., and Riseborough, D. W. (2002). Climate and the limits of permafrost: a zonal approach. *Permafrost and Periglacial Processes*, 13(1), 1–15. <https://doi.org/10.1002/ppp.410>
- Smith, S. L., Riseborough, D. W., and Bonnaventure, P. P. (2015). Eighteen-year record of forest fire effects on ground thermal regimes and permafrost in the central Mackenzie Valley, NWT, Canada. *Permafrost and Periglacial Processes*, 26(4), 289-303. <https://doi.org/10.1002/ppp.1849>
- Smith, S. L., Romanovsky, V. E., Lewkowicz, A. G., Burn, C. R., Allard, M., Clow, G. D., Yoshikawa, K., and Throop, J. (2010). Thermal state of permafrost in North America—a contribution to the International Polar Year. *Permafrost and Periglacial Processes*, 21(2), 117–135. <http://dx.doi.org/10.1002/ppp.690>
- Swanson, D. K. (1996). Susceptibility of permafrost soils to deep thaw after forest fires in interior Alaska, U.S.A., and some ecologic implications. *Arctic and Alpine Research*, 28(2), 217-227.
- Throop, J., Lewkowicz, A.G., and Smith, S.L. (2012). Climate and ground temperature relations at sites across the continuous and discontinuous permafrost zones, northern Canada. *Canadian Journal of Earth Sciences*, 49(8), 865–876. <https://doi.org/10.1139/e11-075>
- Tukey, J. W. (1977). Exploratory data analysis. Addison-Wesley Publishing Co.
- Turetsky, M. R., Donahue, W. F., and Benscoter, B. W. (2011). Experimental drying intensifies burning and carbon losses in a northern peatland. *Nature Communications*, 2, 514. <https://doi.org/10.1038/ncomms1523>
- van Everdingen, R. (2005). Multi-language glossary of permafrost and related ground-ice terms. National Snow and Ice Data Center, Boulder, Colorado, USA.
- Verret, M. (2017). *Distribution, morphology and carbon stock of earth hummocks in the Chuck Creek Trail Valley, northern British Columbia, Canada* [M.Sc. thesis]. University of Ottawa.

- Viereck, L. A., Werdin-Pfisterer, N. R., Adams, P. C., and Yoshikawa, K. (2009). Effect of wildfire and fireline construction on the annual depth of thaw in a black spruce permafrost forest in interior Alaska: a 36-year record of recovery. In D. L. Kane and K. M. Hinkel (Eds.), *Proceedings of the 9th International Conference on Permafrost*, Institute of Northern Engineering, University of Alaska Fairbanks, Volume 2, 1845-1850.
- Walker, X. J., Baltzer, J. L., Cumming, S. G., Day, N. J., Johnstone, J. F., Rogers, B. M., Solvik, K., Turetsky, M. R., and Mack, M. C. (2018a). Soil organic layer combustion in boreal black spruce and jack pine stands of the Northwest Territories, Canada. *International Journal of Wildland Fire*, 27(2), 125-134. <https://doi.org/10.1071/WF17095>
- Walker, X. J., Rogers, B. M., Baltzer, J. L., Cumming, S. R., Day, N. J., Goetz, S., Johnstone, J. F., Schuur, E. A. G., Turetsky, M. R., and Mack, M. C. (2018b). Cross-scale controls on carbon emissions from boreal mega-fires of the Northwest Territories, Canada. *Global Change Biology*, 24(9), 4251-4265. <https://doi.org/10.1111/gcb.14287>
- Wang, X., Thompson, D. K., Marshall, G. A., Tymstra, C., Carr, R., and Flannigan, M. D. (2015). Increasing frequency of extreme fire weather in Canada with climate change. *Climatic Change*, 130, 573-586. <https://doi.org/10.1007/s10584-015-1375-5>
- Way, R. G., Lewkowicz, A. G., and Zhang, Y. (2018). Characteristics and fate of isolated permafrost patches in coastal Labrador, Canada. *The Cryosphere*, 12, 2667-2688. <https://doi.org/10.5194/tc-12-2667-2018>
- Wolfe, S. A., Morse, P. D., Hoeve, T. E., Sladen, W. E., Kokelj, S. V., and Arenson, L. U. (2015). Disequilibrium permafrost conditions on NWT Highway 3. In *Proceedings of the 68th Canadian Geotechnical Conference & 7th Canadian Permafrost Conference*. Canadian Geotechnical Society.
- Wotton, B. M., Flannigan, M. D., and Marshall, G. A. (2017). Potential climate change impacts on fire intensity and key wildfire suppression thresholds in Canada. *Environmental Research Letters*, 12(9), 095003. <https://doi.org/10.1088/1748-9326/aa7e6e>
- Yi, S., McGuire, A. D., Kasischke, E., Harden, J., Manies, K., Mack, M., and Turetsky, M. (2010). A dynamic organic soil biogeochemical model for simulating the effects of wildfire on soil environmental conditions and carbon dynamics of black spruce forests. *Journal of Geophysical Research: Biogeosciences*, 115(G4). <https://doi.org/10.1029/2010JG001302>
- Zhang, Y., Wolfe, S. A., Morse, P. D., Olthof, I., and Fraser, R. H. (2015). Spatiotemporal impacts of wildfire and climate warming on permafrost across a subarctic region, Canada. *Journal of Geophysical Research: Earth Surface*, 120(11), 2338-2356. <https://doi.org/10.1002/2015JF003679>

CHAPTER 6: CONCLUSIONS

6.1 Summary and conclusions

This thesis describes permafrost conditions at more than 80 field sites, and provides knowledge regarding changes to frozen ground over the last half century across the discontinuous zones of northwestern Canada. This includes the impacts of regional climate warming of $\sim 2^{\circ}\text{C}$ since 1945 (Chapter 3) and permafrost response to forest fire (Chapters 3 and 5). This dataset allowed analysis of how permafrost has responded and is responding to disturbance across a wide range of environmental conditions, and permitted direct comparison between burned and unburned sites. Prior to this research, recent observations of permafrost conditions following fire in northwestern Canada were limited in number, especially over the long-term (Chapter 2). This is particularly true for certain variables impacting permafrost response, such as burn severity, and post-fire changes in snow and soil moisture (Chapter 2). The detailed field observations in this thesis augment knowledge of permafrost response to fire in the southern Northwest Territories and northern Alberta. They can be used to understand the response of boreal forest environments elsewhere in Canada and in other countries, and to further modelling efforts to predict permafrost change as the climate warms.

In general, it is concluded that some bodies of near-surface permafrost have completely degraded across the discontinuous zones as a result of ongoing climate change in the region, and additionally, due to fires in the last decade (Chapters 3 and 5). However, permafrost has also persisted at many sites, including disturbed sites as far south as 59.3°N (Chapter 3). This is particularly the case where black spruce (*Picea mariana*) forms the canopy, organic layers are >40 cm thick, and the substrate is fine-grained (Chapters 3 and 5). Less change has occurred at sites with higher gravimetric moisture contents, but this finding must be placed in the context of limited thermokarst development at the sites examined (Chapter 5).

Contrary to the initial hypothesis, permafrost in southerly locations remains resilient to fire if it is protected by a thick organic layer (Chapters 3 and 5) and does not experience ground subsidence and thermokarst (Chapter 5). Similarly, cold permafrost sites at the northern end of the transect appear resilient even when severely burned (Chapter 5). At some sites which burned in the last decade, decreases in resistivity and increases in ground temperature (Chapter 5), as well as deeper frost table depths (Chapters 3 and 5), were observed compared to unburned locations. However, change was mainly in the near-surface, and where observations were

possible of permafrost at depths greater than about 5 m, it was relatively unaffected within the time-frame of five years after fire. Historically, over the long-term, it appears that the majority of sites return to pre-fire conditions, as no persistent changes in active layer thickness were observed at sites which burned as much as 40 years ago (Chapter 3). This is consistent with literature that shows permafrost changes following fire return to pre-fire conditions after several decades (Chapter 2). However, results from Chapter 3 only indicate permafrost resiliency under past climatic conditions, and observations from Chapter 5 are too short-term to conclude whether permafrost will recover to pre-fire conditions. It is also unknown whether permafrost in the region will recover fully following fire in the present or future. Modelling suggests that fires occurring after 1990 cause irreversible permafrost degradation at most forested sites (Zhang et al., 2015). It is therefore expected that frozen ground in the discontinuous zone will become more vulnerable to fire in the future.

Sites where permafrost degraded entirely generally had thin organic layers overlying coarse-grained substrates, and were mostly in the taiga plains ecozone (Chapters 3 and 5). Permafrost in the zone of isolated patches at the southernmost end of the study region has also degraded over the last 50 years, even at sites with organic layers greater than 80 cm thick (Chapter 3). The terrain in this area is now exposed to a climate that cannot sustain permafrost, no matter how robust the ecosystem protection.

Many previous studies have assessed post-fire permafrost degradation by tracking changes in the frost table, usually by probing. This thesis provides information on permafrost conditions at greater depths. Repeated ERT surveys along fixed lines of electrodes revealed localized changes in permafrost to depths of up to 25 m (Chapters 3 and 5). Laboratory measurements of resistivity during soil freezing added to the literature surrounding the interpretation of electrical geophysical surveys and their use in permafrost monitoring (Chapter 4).

Ground ice plays a significant role in the trajectory that permafrost follows after fire (Chapter 2). This thesis focuses on permafrost change following fire at sites with little or no excess ice, conditions which prevail in the study region (Chapter 1; O'Neill et al., 2019; 2020). Supra-permafrost and through-going taliks were observed at some warm sites in the sporadic discontinuous zone, particularly those that experienced severe burning. Observations of ground subsidence and thermokarst, however, were limited, and substantial water inundation and

permafrost thaw occurred at only one monitored site where excess ice was present. Gibson et al. (2018) found extensive permafrost thaw and rapid thermokarst bog development in peatlands following fire in southern NWT, at a rate exceeding that observed in this study. The contrast in findings is likely due to differences between peatlands and forested sites, particularly in relation to hydrological and ground ice conditions. It is clear that predicting the post-fire behaviour of permafrost requires studies across a complete range of extant landscapes (Chapters 2, 3 and 5).

It is concluded that permafrost can be resilient to fire in certain settings, and this underlines the importance of monitoring sites over a broad range of environmental conditions. Post-fire permafrost response is sensitive to individual site characteristics, such as ground ice content, drainage conditions, substrate, and organic layer thicknesses, which dictate the trajectory of change and despite the geographic extent of the transect, largely supersede the regional or latitudinal context. All landscape types and the heterogeneity of their responses need to be considered, therefore, when modelling the post-fire degradation of frozen ground.

6.2 Key contributions

All four of the main thesis objectives, outlined in Chapter 1.2, were met and several key scientific contributions were made regarding the impacts of fire on discontinuous permafrost in northwestern Canada:

1. Synthesis of existing literature on the impacts of fire on permafrost and identification of knowledge gaps;
2. Identification of long-term changes to permafrost at specific sites following climate warming and disturbance by fire;
3. Regional analysis of how permafrost conditions have changed in the discontinuous zones over the last 50 years;
4. Direct observation of short-term permafrost response following fire over a wide range of environmental conditions;
5. Identification of a critical residual organic layer thickness (40 cm) which limits change following fire in the region;
6. Post-fire observations at low ice-content permafrost sites;
7. Novel direct measurements of changes in snow depth following fire;
8. Observations of the impacts of burn severity on permafrost;

9. Quantification of ground subsidence and identification of taliks following fire;
10. Assessment of electrical resistivity tomography as a post-fire monitoring tool;
11. Establishment of permanent permafrost monitoring infrastructure in the southern Northwest Territories, allowing for long-term post-fire measurements.

6.3 Limitations and future research directions

This thesis enhances understanding of permafrost change across a wide range of conditions. However, several limitations remain, and additional research is needed to fully understand the role fire plays on permafrost landscapes. Key knowledge gaps are outlined in Chapter 2.6, but are expanded upon below.

Almost all the sites examined in this thesis were accessed from the highway or road network, which permitted data collection at more than 80 locations between 2015 and 2019. Most of these sites were chosen by Roger Brown in 1962 and were located within 100 m of the road which likely resulted in under-sampling of some types of environment. For example, wetlands make up about 40% of the taiga plains ecozone (Gibson et al., 2018) (although large parts of wetlands do not have permafrost) while fewer than 15% of the sites visited were of this landscape type. Access was particularly important for the monitoring sites because relatively heavy equipment (ERT, batteries, etc.) had to be carried through the burned forest and deployed during each visit. This limited sites to a maximum of 1 km from a road or trail, with the exception of the Lucky Lake sites which were accessed by helicopter. Thermokarst development is more rapid and dramatic in peatland sites (Gibson et al., 2018; Olefeldt et al., 2016), most of which cannot be accessed on foot, so our conclusions are conservative for permafrost change post-fire. In fact, this thesis helps offset the geographic dominance in the literature of studies from high ice-content sites, by mainly examining forested and low-ice content sites (Chapter 2). In the future, statistical or remote sensing techniques could be enlisted to improve site selection and ensure that all landscape types are appropriately represented (e.g. Walker et al., 2018b).

Further work is needed to examine the hypothesis that permafrost near its southernmost margin is ecosystem-protected and so will thaw entirely following fire. No study within the last 10 years has demonstrated this (Chapter 2). In the initial study hypothesis, it was predicted that post-fire impacts to permafrost would follow a latitudinal gradient. However, the impact of fire in the zone of isolated patches could not be addressed as none of the sites visited by Brown in

1962 were burned (Chapter 3). Based on climate, it appears that permafrost at its most southerly margin is ecosystem-protected (Chapter 3). However, sites slightly farther north, where MAAT is approximately -2°C , can be relatively resilient (Chapters 3 and 5). Further direct observations of the post-fire permafrost response at its southern limit is needed to fill this knowledge gap, and validate modelling efforts for future permafrost change in the region.

Chapter 5 partly relied on the assumption that the unburned monitoring sites are representative of pre-fire conditions at the burned sites, and that they act as a baseline for the ongoing impacts of climate change at all sites. This pairing of burned and unburned locations has been used in many other post-fire permafrost monitoring studies (e.g. Nossov et al., 2013; Smith et al., 2015). The limitation in the present study is that the two unburned sites could not capture the wide range of conditions at all burned sites. However, no other unburned sites could be found in 2015, leaving gaps at the two ends of the transect.

The initial choice of monitoring sites was rigorously randomized, but there was the possibility of bias at the local scale: transects were laid out to cover the widest range of burn severities and features such as large ponds were generally avoided. Most permafrost research has been conducted in this way, but future studies quantifying regional permafrost extent or change should consider the distribution of landscape types when sampling. Quantitative regional upscaling from site-specific measurements requires improved remote sensing techniques to link surface characteristics to post-fire permafrost degradation.

A limitation that applies to all field studies was the lack of control over climate. Although the long-term trend is towards warming in the region, cooling took place during the period of record. These contrary influences presented challenges for data interpretation.

There are several limitations associated with ERT (Hauck et al., 2013) despite its widening use in permafrost studies. Although a great deal of work went into validating results using field and laboratory techniques (Chapter 4), ERT interpretation was still partly subjective. Thresholds between frozen and unfrozen ground were not calculated at the majority of sites, and these locations relied on frost probing as validation. This methodology has been used elsewhere to validate ERT results (e.g. Way et al., 2018), but is more difficult for coarse-grained substrates. Interpretation was difficult at sites with dry sand or bedrock, which have modelled resistivities similar to permafrost. Repeated ERT helps overcome some of these concerns and permafrost

presence was still able to be established at all sites using a combination of methods, including temperature monitoring and frost probing.

Frost probing, while dating from decades before ERT, also had some limitations. The frost probe employed throughout this thesis was 110 cm long, and there was no direct way of determining whether a frost table was present if this depth was reached without refusal. To overcome this limitation, a combination of ERT and temperature gradients (Table B2) were used, but neither are direct observations. There were general issues with probing at sites with coarse substrates, as clasts can be misinterpreted as frozen ground. In those instances, real-time temperature measurements using analog data loggers were utilized to aid in validation of probing. Similarly, frozen ground coring was not possible at sites with coarse substrates. Pits were dug at these locations, but a comprehensive assessment of the deeper soil profile was not possible. Cost- and time-effective techniques for sampling permafrost in coarse-grained substrates are needed.

This thesis added to a body of literature that shows the impacts of increasing burn severity on permafrost at specific sites (Jafarov et al., 2013; Jiang et al., 2015; Li et al., 2019). However, more work is needed to upscale these measurements to the landscape level, and link them to expected permafrost response in the future. For example, Zhang et al. (2015) addressed permafrost change over the next century with climate warming and fire at a regional scale, but did not incorporate burn severity. Data from Chapter 5 shows that burn severity at the site level is heterogeneous, supporting other literature that shows large parts of mapped fire areas may remain unburned (Fraser et al., 2003; Kaschiske et al., 2010). Despite these observations, in Chapter 3 an assumption that the entire fire scar polygon burned was necessary. Site-specific observations were made to assess burned locations, but exact validation was not always possible. In the future, dendrochronology could be used to assess time since the last fire (e.g. Walker et al., 2018a).

This thesis provides the first direct measurements of changes in snow depth following fire in permafrost environments. However, more observations are desirable to show how snowpack conditions change over time following fire depending on burn severity, and the coupled impacts on ground temperatures. Chapter 5 describes only the average SDD and maximum snow depth at a particular site, and more nuanced seasonal patterns are needed. This is particularly important because fire has been shown to cause changes in snowmelt date in non-

permafrost settings (Burles and Boon, 2011; Gleason et al., 2013), and this will impact permafrost through the nival offset (Jorgenson et al., 2010). Further, it is unclear what happens at non-forested wetland sites, where post-fire changes in snow patterns may be negligible. Modelling efforts have incorporated some of these effects of snow across different landscapes (Jafarov et al., 2013; Zhang et al., 2015) but more direct measurements are needed for validation.

In general, further analysis is needed to link empirical studies showing permafrost change following fire with modelling efforts. Results from Chapter 3 provide important long-term validation for the latter, and data from Chapter 5 can be incorporated into models to more accurately predict how permafrost responds to fire. However, models currently do not fully integrate thermokarst development, so likely underestimate permafrost thaw following fire. Models point to the resiliency of lowland boreal forest permafrost following fire (Jafarov et al., 2013; Zhang et al., 2015), but lack the observations of rapid thermokarst development found in some empirical studies (Brown et al., 2015; Gibson et al., 2018). Data from this thesis support both trajectories for lowland permafrost post-fire, but it remains unclear exactly what thresholds trigger thermokarst initiation in lowland environments. Future work in this area is needed to fully understand the impacts of fire on permafrost landscapes.

Permafrost response to fire is a combination of complex relationships between the burn itself, climate, vegetation, soils, hydrology and ground thermal state. This means that no single discipline can fully evaluate the impacts of fire on permafrost, and it is necessary to bridge results from the ecological, geocryological, biogeochemical, remote sensing and modelling communities. For example, this study lacked robust information to link post-fire ecological and permafrost changes. Particularly useful would have been data to establish post-fire successional trajectories, as state changes due to severe burning have also been observed elsewhere (Johnstone et al., 2010), but the relationship to permafrost requires further study. Overall, more work is needed to link vegetation recovery, organic layer regeneration, and permafrost change in a warming climate.

6.4 References

- Brown, D. R. N., Jorgenson, M. T., Douglas, T. A., Vladimir, R. E., Kielland, K., Hiemstra, C., Euskirchen, E. S., and Ruess, R. W. (2015). Interactive effects of wildfire and climate on permafrost degradation in Alaskan lowland forests. *Journal of Geophysical Research - Biogeosciences*, 120(8), 1619-1637. <https://doi.org/10.1002/2015JG003033>
- Burles, K., and Boon, S. (2011). Snowmelt energy balance in a burned forest plot, Crowsnest Pass, Alberta, Canada. *Hydrological Processes*, 25(19), 3012-3029. <https://doi.org/10.1002/hyp.8067>
- Fraser, R. H., Hall, R. J., Landry, R., Lynham, T. J., Raymond, D. A., Lee, B. S., and Li, Z. (2004). Validation and calibration of Canada-wide coarse-resolution satellite burned-area maps. *Photogrammetric Engineering & Remote Sensing*, 70, 451-460. <https://doi.org/10.14358/PERS.70.4.451>
- Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., and Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nature Communications*, 9, 3041. <https://doi.org/10.1038/s41467-018-05457-1>
- Gleason, K. E., Nolin, A. W., and Roth, T. R. (2013). Charred forests increase snowmelt: effects of burned woody debris and incoming solar radiation on snow ablation. *Geophysical Research Letters*, 40(17), 4654-4661. <https://doi.org/10.1002/grl.50896>
- Hauck, C. (2013). New concepts in geophysical surveying and data interpretation for permafrost terrain. *Permafrost and Periglacial Processes*, 24(2), 131-137. <https://doi.org/10.1002/ppp.1774>
- Jafarov, E. E., Romanovsky, V. E., Genet, H., McGuire, A. D., and Marchenko, S. S. (2013). The effects of fire on the thermal stability of permafrost in lowland and upland black spruce forests of interior Alaska in a changing climate. *Environmental Research Letters*, 8(3), 035030. <http://dx.doi.org/10.1088/1748-9326/8/3/035030>
- Jiang, Y., Rocha, A. V., O'Donnell, J. A., Drysdale, J., Rastetter, E. B., Shaver, G. R., and Zhuang, Q. (2015). Contrasting soil thermal responses to fire in Alaskan tundra and boreal forest. *Journal of Geophysical Research: Earth Surface*, 120(2), 363-378. <http://dx.doi.org/10.1002/2014JF003180>
- Johnstone, J. F., Hollingsworth, T. N., Chapin, F. S., and Mack, M. C. (2010). Changes in fire regime break the legacy lock on successional trajectories in Alaskan boreal forest. *Global Change Biology*, 16(4), 1281-1295. <https://doi.org/10.1111/j.1365-2486.2009.02051.x>
- Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., O'Donnell, J., Schuur, E. A. G., Kanevskiy, M., and Marchenko, S. (2010). Resilience and vulnerability of permafrost to climate change. *Canadian Journal of Forest Research*, 40(7), 1219-1236. <https://doi.org/10.1139/X10-060>

- Kasischke, E. S., Verbyla, D. L., Rupp, T. S., McGuire, A. D., Murphy, K. A., Jandt, R., Barnes, J. L., Hoy, E. E., Duffy, P. A., Calef, M., and Turetsky, M. R. (2010). Alaska's changing fire regime - implications for the vulnerability of its boreal forests. *Canadian Journal of Forest Research*, 40(7), 1313-1324. <https://doi.org/10.1139/X10-098>
- Kwong, Y. J., and Gan, T. Y. (1994). Northward migration of permafrost along the Mackenzie Highway and climatic warming. *Climatic Change*, 26(4), 399-419. <https://doi.org/10.1007/BF01094404>
- Li, X., Jin, H., He, R., Huang, Y., Wang, H., Luo, D., Jin, X., Lü, L., Wang, L., Li, W., Wei, C., Chang, X., Yang, S., and Yu, S. (2019). Effects of forest fires on the permafrost environment in the northern Da Xing'anling (Hinggan) mountains, Northeast China. *Permafrost and Periglacial Processes*, 30(3), 163-177. <https://doi.org/10.1002/ppp.2001>
- Nossov, D. R., Jorgenson, M. T., Kielland, K., and Kanevskiy, M. Z. (2013). Edaphic and microclimatic controls over permafrost response to fire in interior Alaska. *Environmental Research Letters*, 8(3), 035013. <https://doi.org/10.1088/1748-9326/8/3/035013>
- Olefeldt, D., Goswami, S., Grosse, G., Hayes, D., Hugelius, G., Kuhry, P., McGuire, A. D., Romanovsky, V. E., Sannel, A. B. K., Schuur, E. A. G., Turetsky, M. R. (2016). Circumpolar distribution and carbon storage of thermokarst landscapes. *Nature Communications*, 7, 13043. <https://doi.org/10.1038/ncomms13043>
- O'Neill, H. B., Wolfe, S. A., and Duchesne, C. (2019). New ground ice maps for Canada using a paleogeographic modelling approach. *Cryosphere*, 13, 753–773. <https://doi.org/10.5194/tc-13-753-2019>
- O'Neill, H. B., Wolfe, S. A., and Duchesne, C. (2020). *Ground ice map of Canada v.1*. Geological Survey of Canada, Open File 8713. In press.
- Smith, S. L., Riseborough, D. W., and Bonnaventure, P. P. (2015). Eighteen-year record of forest fire effects on ground thermal regimes and permafrost in the central Mackenzie Valley, NWT, Canada. *Permafrost and Periglacial Processes*, 26(4), 289-303. <https://doi.org/10.1002/ppp.1849>
- Walker, X. J., Baltzer, J. L., Cumming, S. G., Day, N. J., Johnstone, J. F., Rogers, B. M., Solvik, K., Turetsky, M. R., and Mack, M. C. (2018a). Soil organic layer combustion in boreal black spruce and jack pine stands of the Northwest Territories, Canada. *International Journal of Wildland Fire*, 27(2), 125-134. <https://doi.org/10.1071/WF17095>
- Walker, X. J., Rogers, B. M., Baltzer, J. L., Cumming, S. R., Day, N. J., Goetz, S., Johnstone, J. F., Schuur, E. A. G., Turetsky, M. R., and Mack, M. C. (2018b). Cross-scale controls on carbon emissions from boreal mega-fires of the Northwest Territories, Canada. *Global Change Biology*, 24(9), 4251-4265. <https://doi.org/10.1111/gcb.14287>

- Way, R. G., Lewkowicz, A. G., and Zhang, Y. (2018). Characteristics and fate of isolated permafrost patches in coastal Labrador, Canada. *The Cryosphere*, 12, 2667-2688. <https://doi.org/10.5194/tc-12-2667-2018>
- Zhang, Y., Wolfe, S. A., Morse, P. D., Olthof, I., and Fraser, R. H. (2015). Spatiotemporal impacts of wildfire and climate warming on permafrost across a subarctic region, Canada. *Journal of Geophysical Research: Earth Surface*, 120(11), 2338-2356. <https://doi.org/10.1002/2015JF003679>

THESIS BIBLIOGRAPHY

- Aaltonen, H., Palviainen, M., Zhou, X., Köster, E., Berninger, F., Pumpanen, J., and Köster, K. (2019). Temperature sensitivity of soil organic matter decomposition after forest fire in Canadian permafrost region. *Journal of Environmental Management*, 241, 637-644. <https://doi.org/10.1016/j.jenvman.2019.02.130>
- ACIA. (2005). *Arctic Climate Impact Assessment*. Cambridge University Press. 1020 pp.
- Alberta Government. (2018). *Historical Fire Database*. <http://wildfire.alberta.ca/resources/historical-data/historical-wildfire-database.aspx>
- Allison, S. D., and Treseder, K. K. (2011). Climate change feedbacks to microbial decomposition in boreal soils. *Fungal Ecology*, 4(6), 362-374. <https://doi.org/10.1016/j.funeco.2011.01.003>
- Andersland, O., and Ladanyi, B. (2004). *Frozen Ground Engineering*. 2nd ed. Chichester, UK: John Wiley & Sons.
- AMAP. (2017). *Snow, Water, Ice and Permafrost in the Arctic (SWIPA)*. Oslo, Norway: Arctic Monitoring and Assessment Programme.
- Amiro, B. D. (1999). BOREAS flight measurements of forest-fire effects on carbon dioxide and energy fluxes. *Agricultural and forest meteorology*, 96(4), 199-208. [https://doi.org/10.1016/s0168-1923\(99\)00050-7](https://doi.org/10.1016/s0168-1923(99)00050-7)
- Amiro, B. D. (2001). Paired-tower measurements of carbon and energy fluxes following disturbance in the boreal forest. *Global Change Biology*, 7(3), 253-268. <https://doi.org/10.1046/j.1365-2486.2001.00398.x>
- Amiro, B. D., Orchansky, A. L., Barr, A. G., Black, T. A., Chambers, S. D., Chapin III, F. S., Goulden, M. L., Litvak, M., Liu, H. P., McCaughey, J. H., McMillan, A., Randerson, J. T. (2006). The effect of post-fire stand age on the boreal forest energy balance. *Agricultural and Forest Meteorology*, 140(1-4), 41-50. <https://doi-org.proxy.bib.uottawa.ca/10.1016/j.agrformet.2006.02.014>
- Anderson, L., Edwards, M., Shapley, M. D., Finney, B. P., and Langdon, C. (2019). Holocene thermokarst lake dynamics in northern Interior Alaska: the interplay of climate, fire, and subsurface hydrology. *Frontiers in Earth Science*, 7(53), 1-22. <https://doi.org/10.3389/feart.2019.00053>
- Balshi, M. S., McGuire, A. D., Duffy, P., Flannigan, M., Walsh, J., Melillo, J. (2009). Assessing the response of area burned to changing climate in western boreal North America using a Multivariate Adaptive Regression Splines (MARS) approach. *Global Change Biology*, 15(3), 578-600. <https://doi.org/10.1111/j.1365-2486.2008.01679.x>
- Baltzer, J. L., Veness, T., Chasmer, L. E., Sniderhan, A. E., and Quinton, W. L. (2014). Forests on thawing permafrost: fragmentation, edge effects, and net forest loss. *Global Change Biology*, 20(3), 824-834. <https://doi.org/10.1111/gcb.12349>

- Bai, Q., Zhou, M., Qi, Q., Meng, R., Cui, X., and Wen, D. (2013). The changes of the frozen soil temperature and humidity of the forest fire forest in the Great Hinggan Mountains. *Journal of Arid Land Resources and Environment*, 5(27), 69-75.
- Beck, P. S. A., Goetz, S. J., Mack, M. C., Alexander, H. D., Jin, Y., Randerson, J. T., Loranty, M. M. (2011). The impacts and implications of an intensifying fire regime on Alaskan boreal forest composition and albedo. *Global Change Biology*, 17(9), 2853–2866. <https://doi.org/10.1111/j.1365-2486.2011.02412.x>
- Beilman, D. W., and Robinson, S. D. (2003). Recent permafrost melt along a climatic gradient in western Canada: large-scale peat mapping. In M. Phillips, S. M. Springman, L. U. Arenson, A. A. Balkema (Eds.), *Proceedings of the 8th International Conference on Permafrost* (pp. 61-65). Zurich, Switzerland.
- Beltaos, S. and Prowse, T. (2009). River-ice hydrology in a shrinking cryosphere. *Hydrological Processes*, 23(1), 122-144. <https://doi.org/10.1002/hyp.7165>
- Bem, D. J. (1995). Writing a review article for psychological bulletin. *Psychological Bulletin*, 118(2), 172–177.
- Benn, D., and Evans, D. J. (2010). Chapter 2: Snow, Ice and Climate. In *Glaciers and Glaciation*. Second Edition. Hodder Education. London, UK.
- Benscoter, B. W., Thompson, D. K., Waddington, J. M., Flannigan, M. D., Wotton, B. M., De Groot, W. J., and Turetsky, M. R. (2011). Interactive effects of vegetation, soil moisture and bulk density on depth of burning of thick organic soils. *International Journal of Wildland Fire*, 20(3), 418-429. <https://doi.org/10.1071/WF08183>
- Benson, B. J., Magnuson, J. J., Jensen, O. P., Card, V. M., Hodgkins, G., Korhonen, J., Livingstone, D. M., Stewart, K. M., Weyhenmeyer, G. A. and Granin, N. G. (2012). Extreme events, trends, and variability in Northern Hemisphere lake-ice phenology (1855–2005). *Climatic Change*, 112, 299-323. <https://doi.org/10.1007/s10584-011-0212-8>
- Boby, L. A., Schuur, E. A., Mack, M. C., Verbyla, D., and Johnstone, J. F. (2010). Quantifying fire severity, carbon, and nitrogen emissions in Alaska's boreal forest. *Ecological Applications*, 20, 1633-1647. <https://doi.org/10.1890/08-2295.1>
- Bret-Harte, M. S., Mack, M. C., Shaver, G. R., Huebner, D. C., Johnston, M., Mojica, C. A., Pizano, C., and Reiskind, J. A. (2013). The response of Arctic vegetation and soils following an unusually severe tundra fire. *Philosophical Transactions of the Royal Society B: Biological Sciences*, 368(1624), 20120490. <https://doi.org/10.1098/rstb.2012.0490>
- Brown, J., Hinkel, K. M., and Nelson, F. E. (2000) The Circumpolar Active Layer Monitoring (CALM) program: Research designs and initial results. *Polar Geography*, 24(3), 165–258.

- Brown, D. R. N., Jorgenson, M. T., Douglas, T. A., Vladimir, R. E., Kielland, K., Hiemstra, C., Euskirchen, E. S., and Ruess, R. W. (2015). Interactive effects of wildfire and climate on permafrost degradation in Alaskan lowland forests. *Journal of Geophysical Research - Biogeosciences*, 120(8), 1619-1637. <https://doi.org/10.1002/2015JG003033>
- Brown, R. J. E. (1964). *Permafrost investigations on the Mackenzie Highway in Alberta and Mackenzie district*. Ottawa, Canada, National Research Council Technical Paper No. 175.
- Bulygina, O. N., Razuvaev, V. N., and Korshunova, N. N. (2009). Changes in snow cover over Northern Eurasia in the last few decades. *Environmental Research Letters*, 4(4), 045026. <https://doi.org/10.1088/1748-9326/4/4/045026>
- Burd, K., Tank, S. E., Dion, N., Quinon, W. L., Spence, C., Tanentzap, A. J., and Olefeldt, D. (2018). Seasonal shifts in export of DOC and nutrients from burned and unburned peatland-rich catchments, Northwest Territories, Canada. *Hydrology and Earth System Science*, 22(8), 4455-4472. <https://doi.org/10.5194/hess-22-4455-2018>
- Burles, K., and Boon, S. (2011). Snowmelt energy balance in a burned forest plot, Crowsnest Pass, Alberta, Canada. *Hydrological Processes*, 25(19), 3012-3029. <https://doi.org/10.1002/hyp.8067>
- Burn, C. R. (1998). The response (1958-1997) of permafrost and near-surface ground temperatures to forest fire, Takhini River Valley, southern Yukon Territory. *Canadian Journal of Earth Science*, 35(2), 184-199. <https://doi.org/10.1139/e97-105>
- Burn, C. R. and Kokelj, S. V. (2009). The environment and permafrost of the Mackenzie Delta area. *Permafrost and Periglacial Processes*, 20(2), 83-105. <https://doi.org/10.1002/ppp.655>
- Bush, E., and Lemmen, D. S., (eds.). (2019). *Canada's Changing Climate Report*. Government of Canada, Ottawa, ON. <https://changingclimate.ca/CCCR2019/>
- Camill, P. (2005). Permafrost thaw accelerates in boreal peatlands during late-20th century climate warming. *Climatic Change*, 68, 135-152. <https://doi.org/10.1007/s10584-005-4785-y>
- Camill, P., and Clark, J. S. (1998). Climate change disequilibrium of boreal permafrost peatlands caused by local processes. *American Naturalist*, 151(3), 207-222. <https://doi.org/10.1086/286112>
- Canadian Interagency Forest Fire Center. (2014). *Situation Report – Sep 22, 2014*. <http://www.cifffc.ca/firewire/current.php?lang=en&date=20140922>
- Chambers, S. D., and Chapin III, F. S. (2002). Fire effects on surface-atmosphere energy exchange in Alaskan black spruce ecosystems: implications for feedbacks to regional climate. *Journal of Geophysical Research*, 107(D1), FFR 1-1 - FFR 1-17. <https://doi.org/10.1029/2001JD000530>

- Chambers S. D., Beringer, J., Randerson, J. T., Chapin III, F. S. (2005). Fire effects on net radiation and energy partitioning: contrasting responses of tundra and boreal forest ecosystems. *Journal of Geophysical Research*, 110(D9). <https://doi.org/10.1029/2004JD005299>
- Chipman, M. L., Hudspith, V., Higuera, P. E., Duffy, P. A., Kelly, R., Oswald, W. W., and Hu, F. S. (2015). Spatiotemporal patterns of tundra fires: Late-Quaternary charcoal records from Alaska. *Biogeosciences*, 12, 4017-4027. <https://doi.org/10.5194/bg-12-4017-2015>
- Connon, R., Devoie, É., Hayashi, M., Veness, T., and Quinton, W. (2018). The influence of shallow taliks on permafrost thaw and active layer dynamics in subarctic Canada. *Journal of Geophysical Research: Earth Surface*, 123(2), 281-297. <https://doi.org/10.1002/2017JF004469>
- Connon, R. F., Quinton, W. L., Craig, J. R., and Hayashi, M. (2014). Changing hydrologic connectivity due to permafrost thaw in the lower Liard River valley, NWT, Canada. *Hydrological Processes*, 28(14), 4163-4178. <https://doi.org/10.1002/hyp.10206>
- Day, N. J., Dunfield, K. E., Johnstone, J. F., Mack, M. C., Turetsky, M. R., Walker, X. J., White, A. L., and Baltzer, J. L. (2019). Wildfire severity reduces richness and alters composition of soil fungal communities in boreal forests of western Canada. *Global Change Biology*, 25(7), 2310-2324. <https://doi.org/10.1111/gcb.14641>
- de Groot, W. J., Flannigan, M. D., and Cantin, A. S. (2013). Climate change impacts on future boreal fire regimes. *Forest Ecology and Management*, 294, 35-44. <https://doi.org/10.1016/j.foreco.2012.09.027>
- Dieleman, C. M., Branfireun, B. A., McLaughlin, J. W., and Lindo, Z. (2015). Climate change drives a shift in peatland ecosystem plant community: Implications for ecosystem function and stability. *Global Change Biology*, 21(1), 388-395. <https://doi.org/10.1111/gcb.12643>
- Dillon, G. K., Holden, Z. A., Morgan, P., Crimmins, M. A., Heyerdahl, E. K., and Luce, C. H. (2011). Both topography and climate affected forest and woodland burn severity in two regions of the western US, 1984 to 2006. *Ecosphere*, 2(12), 1-33. <https://doi.org/10.1890/ES11-00271.1>
- Doetsch, J., Ingeman-Nielsen, T., Christiansen, A. V., Fiandaca, G., Auken, E., and Elberling, B. (2015). Direct current (DC) resistivity and induced polarization (IP) monitoring of active layer dynamics at high temporal resolution. *Cold Regions Science and Technology*, 119, 16-28. <https://doi.org/10.1016/j.coldregions.2015.07.002>
- Doré, G., Niu, F., and Brooks, H. (2016). Adaptation methods for transportation infrastructure built on degrading permafrost. *Permafrost and Periglacial Processes*, 27(4), 352-364. <https://doi.org/10.1002/ppp.1919>

- Douglas, T. A., Jones, M. C., Hiemstra, C. A., and Arnold, J. R. (2014). Sources and sinks of carbon in boreal ecosystems of interior Alaska: A review. *Elementa Science of the Anthropocene*, 2, 000032. <http://doi.org/10.12952/journal.elementa.000032>
- Douglas, T. A., Jorgenson, M. T., Brown, D. R., Campbell, S. W., Hiemstra, C. A., Saari, S. P., Bjella, K., and Liljedahl, A. K. (2016). Degrading permafrost mapped with electrical resistivity tomography, airborne imagery and LiDAR, and seasonal thaw measurements. *Geophysics*, 81(1), 1JF-Z7. <https://doi.org/10.1190/geo2015-0149.1>
- Douglas, T. A., Jorgenson, M. T., Kanevskiy, M. Z., Romanovsky, V. E., Shur, Y., and Yoshikawa, K. (2008). Permafrost dynamics at the Fairbanks Permafrost Experimental Station near Fairbanks, Alaska. In D. Kane, and K. Hinkle (Eds.), *Proceedings of the Ninth International Conference on Permafrost* (pp. 373-378). University of Alaska Fairbanks.
- Dyrness, C. T. (1982). *Control of depth to permafrost and soil temperature by the forest floor in black spruce/feathermoss communities*. USDA Forest Service Pacific Northwest Forest and Range Experiment Station. <https://www.fs.usda.gov/treearch/pubs/7656>
- Ecological Stratification Working Group. (1996). *A National Ecological Framework for Canada*. Centre for Land and Biological Resources Research, Research Branch, Agriculture and Agri-Food Canada.
- Elmendorf, S. C., Henry, G. H. R., Hollister, R. D., Björk, R. G., Boulanger-Lapointe, N., Cooper, E. J., Cornelissen, J. H. C., Day, T. A., Dorrepaal, E., Elumeeva, T. G., Gill, M., Gould, W. A., Harte, J., Hik, D. S., Hofgaard, A., Johnson, D. R., Johnston, J. F., Jónsdóttir, I. S., Jorgenson, J. C., ... Wipf, S. (2012). Plot-scale evidence of tundra vegetation change and links to recent summer warming. *Nature Climate Change*, 2, 453-457. <https://doi.org/10.1038/nclimate1465>
- Englefield, P. G. C. (1995). *Rate of permafrost degradation in peatlands* [M.Sc. thesis]. University of Alberta. <https://doi.org/10.7939/R3833N81F>
- Environment Canada. (2018). *Historical climate data*. http://climate.weather.gc.ca/historical_data/search_historic_data_e.html
- Environment Canada. (2019). *Historical climate data*. http://climate.weather.gc.ca/historical_data/search_historic_data_e.html.
- Farouki, O. T. (1981). *Thermal properties of soils*. Cold Regions Research and Engineering Lab.
- Fisher, J. P., Estop-Aragonés, C., Thierry, A., Charman, D. J., Wolfe, S. A., Hartley, I. P., Murton, J. B., Williams, M., and Phoenix, G. K. (2016). The influence of vegetation and soil characteristics on active-layer thickness of permafrost soils in boreal forest. *Global Change Biology*, 22, 3127-3140. <https://doi.org/10.1111/gcb.13248>

- Fisher, D. A., Lacelle, D., and Pollard, W. (2019). A model of unfrozen water content and its transport in icy permafrost soils: Effects on ground ice content and permafrost stability. *Permafrost and Periglacial Processes*, 31(1), 184-199. <https://doi.org/10.1002/ppp.2031>
- Flannigan, M. D., Logan, K. A., Amiro, B. D., Skinner, W. R., and Stocks, B. J. (2005). Future area burned in Canada. *Climatic Change*, 72, 1–16. <https://doi.org/10.1007/s10584-005-5935-y>
- Fortier, R., LeBlanc, A. M., Allard, M., Buteau, S., and Calmels, F. (2008). Internal structure and conditions of permafrost mounds at Umiujaq in Nunavik, Canada, inferred from field investigation and electrical resistivity tomography. *Canadian Journal of Earth Science*, 45(3), 367–387. <https://doi.org/10.1139/E08-004>
- Fraser, R. H., Hall, R. J., Landry, R., Lynham, T. J., Raymond, D. A., Lee, B. S., and Li, Z. (2004). Validation and calibration of Canada-wide coarse-resolution satellite burned-area maps. *Photogrammetric Engineering & Remote Sensing*, 70, 451–460. <https://doi.org/10.14358/PERS.70.4.451>
- French, H. M. (2018). *The Periglacial Environment* (4th ed.). John Wiley & Sons, Ltd.
- Genet, H., McGuire, A. D., Barrett, K., Breen, A., Euskirchen, E. S., Johnstone, J. F., Kasischke, E. S., Melvin, A. M., Bennett, A., Mack, M. C., Rupp, T. S., Schuur, A. E. G., Turetsky, M. R., and Yuan, F. (2013). Modelling the effects of fire severity and climate warming on active layer thickness and soil carbon storage of black spruce forests across the landscape in interior Alaska. *Environmental Research Letters*, 8, 045016. <http://dx.doi.org/10.1088/1748-9326/8/4/045016>
- Gibson, C. M., Chasmer, L. E., Thompson, D. K., Quinton, W. L., Flannigan, M. D., and Olefeldt, D. (2018). Wildfire as a major driver of recent permafrost thaw in boreal peatlands. *Nature Communications*, 9, 3041. <https://doi.org/10.1038/s41467-018-05457-1>
- Gleason, K. E., Nolin, A. W., and Roth, T. R. (2013). Charred forests increase snowmelt: effects of burned woody debris and incoming solar radiation on snow ablation. *Geophysical Research Letters*, 40(17), 4654-4661. <https://doi.org/10.1002/grl.50896>
- GNWT. (2018). *GNWT Dataset Download Categories – Forest Management – Fire History*. Northwest Territories Centre for Geomatics. <http://www.geomatics.gov.nt.ca/dldsoptions.aspx>
- Gordon, J., Quinton, W., Branfireun, B. A., and Olefeldt, D. (2016). Mercury and methylmercury biogeochemistry in a thawing permafrost wetland complex, Northwest Territories, Canada. *Hydrological Processes*, 30(20), 3627-3638. <https://doi.org/10.1002/hyp.10911>
- Gu, H., Jin, J., Cheng, X., Wang, E., Zhou, Y., and Chai, Y. (2010). The long-term impacts on chemical properties of *Larix gmelini* forest on the northern slope of Greater Hinggan Mountains from a forest fire of varying fire intensity. *Journal of Natural Resources*, 25(7), 1114-1121 (in Chinese).

- Guo, A., Guo, Y., and Cui, X. (2011). Effects of different intensities of fire disturbances on soil nutrients in a *Pinus massoniana* forest in the Greater Xing'an Mountain. *Journal of Northeast Forestry University*, 39(5), 69-71.
- Halsey, L. A., Vitt, D. H., and Zoltai, S. C. (1995). Disequilibrium response of permafrost in Boreal Continental Western Canada to climate change. *Climatic Change*, 30, 57-73. <https://doi.org/10.1007/BF01093225>
- Hanes, C. C., Wang, X., Jain, P., Parisien, M., Little, J. M., and Flannigan, M. D. (2019). Fire-regime changes in Canada over the last half century. *Canadian Journal of Forest Research*, 49(3), 256-269. <https://doi.org/10.1139/cjfr-2018-0293>
- Hauck, C. (2013). New concepts in geophysical surveying and data interpretation for permafrost terrain. *Permafrost and Periglacial Processes*, 24(2), 131-137. <https://doi.org/10.1002/ppp.1774>
- Heginbottom, J. A., Dubreuil, M. A., and Harker, P. A. (1995). Canada – Permafrost. *National Atlas of Canada* (5th ed.). Natural Resources Canada.
- Heiri, O., Lotter, A.F., and Lemcke, G. (2001). Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. *Journal of Paleolimnology*, 25(1), 101-110. <https://doi.org/10.1023/A:1008119611481>
- Hilbich, C., Fuss, C., and Hauck, C. (2011). Automated time-lapse ERT for improved process analysis and monitoring of frozen ground. *Permafrost and Periglacial Processes*, 22(4), 306-319. <https://doi.org/10.1002/ppp.732>
- Hilbich, C., Marescot, L., Hauck, C., Loke, M. H., and Mäusbacher, R. (2009). Applicability of electrical resistivity tomography monitoring to coarse blocky and ice-rich permafrost landforms. *Permafrost and Periglacial Processes*, 20(3), 269-284. <https://doi.org/10.1002/ppp.652>
- Hinchey, G. (2015, May 6). 2014 N.W.T. fire season report: What you need to know. *CBC North*. <https://www.cbc.ca/news/canada/north/2014-n-w-t-fire-season-report-what-you-need-to-know-1.3061930#:~:text=In%202014%2C%20385%20fires%20burned,million%20hectares%20of%20forest%20land.&text=Despite%20the%20scale%20of%20last,no%20serious%20injuries%20or%20deaths>.
- Hinzman, L. D., Fukuda, M., Sandberg, D. V., Chapin, F. S., and Dash, D. (2003). FROSTFIRE: an experimental approach to predicting the climate feedbacks from the changing boreal fire regime. *Journal of Geophysical Research Atmospheres*, 108(D1), FFR 9-1-FFR 9-6. <https://doi.org/10.1029/2001JD000415>
- Hoekstra, P., Sellmann, P. V., and Delaney, A. J. (1974). *Airborne resistivity mapping of permafrost near Fairbanks, Alaska*. Cold Regions Research and Engineering Laboratory, Report 324.

- Holloway, J., and Lewkowicz, A. G. (2019). Field and laboratory investigation of electrical resistivity-temperature relationships, southern Northwest Territories. In J.-P. Bilodeau, D.F. Nadeau, D. Fortier, and D. Conciatori (eds.), *Cold Regions Engineering 2019: Proceedings of the 18th International Conference on Cold Regions Engineering and 8th Canadian Permafrost Conference* (pp. 64-72). American Society of Civil Engineers, Reston, Virginia. <https://doi.org/10.1061/9780784482599>
- Holloway, J. E., and Lewkowicz, A. G. (2020). Persistence of permafrost following 55 years of climate warming and forest fire in the discontinuous zone, Canada. *Permafrost and Periglacial Processes*, 31(1), 85-96. <https://doi.org/10.1002/ppp.2017>
- Holloway, J. E., Lewkowicz, A. G., Douglas, T. A., Li, X., Turetsky, M. R., Baltzer, J. L., Jin, H. (2020). The impact of wildfire on permafrost landscapes: a review of recent advances and future prospects. *Permafrost and Periglacial Processes*, 31(3), 371-382. <https://doi.org/10.1002/ppp.2048>
- Hoy, E. E., Turetsky, M. R., and Kasischke, E. S. (2016). More frequent burning increases vulnerability of Alaskan boreal black spruce forests. *Environmental Research Letters*, 11, 095001. <http://dx.doi.org/10.1088/1748-9326/11/9/095001>
- Hu, H., Wei, S., and Sun, L. (2012). Estimating carbon emissions from forest fires during 2001 to 2010 in Daxing' anling Mountain. *Shengtai Xuebao/Acta Ecologica Sinica*, 32(17), 5373-5386. <http://dx.doi.org/10.5846/stxb201109091324>
- Hugelius, G., Strauss, J., Zubrzycki, S., Harden, J.W., Schuur, E. A. G., Ping, C.-L., Schirrmeister, L., Grosse, G., Michaelson, G. J., Koven, C. D., O'Donnell, J. A., Elberling, B., Mishra, U., Camill, P., Yu, Z., and Palmtag, J. (2014). Estimated stocks of circumpolar permafrost carbon with quantified uncertainty ranges and identified data gaps. *Biogeosciences*, 11, 6573-6593. <https://doi.org/10.5194/bg-11-6573-2014>
- Huscroft, C. A., Lipovsky, P. S., and Bond, J. D. (2004). *A regional characterization of landslides in the Alaska Highway corridor, Yukon*. Whitehorse, Canada: Yukon Geological Survey, Open File 18.
- IPCC. (2013). Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. T. F. Stocker, D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex, and P. M. Midgley (Eds.). Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp. doi:10.1017/CBO9781107415324.
- Jafarov, E. E., Romanovsky, V. E., Genet, H., McGuire, A. D., and Marchenko, S. S. (2013). The effects of fire on the thermal stability of permafrost in lowland and upland black spruce forests of interior Alaska in a changing climate. *Environmental Research Letters*, 8(3), 035030. <http://dx.doi.org/10.1088/1748-9326/8/3/035030>

- James, M., Lewkowicz, A. G. Smith, S. L., and Miceli C. M. (2013). Multi-decadal degradation and persistence of permafrost in the Alaska Highway corridor, northwest Canada. *Environmental Research Letters* 8(4): 045013. <http://dx.doi.org/10.1088/1748-9326/8/4/045013>.
- Jiang, Y., Rocha, A. V., O'Donnell, J. A., Drysdale, J., Rastetter, E. B., Shaver, G. R., and Zhuang, Q. (2015). Contrasting soil thermal responses to fire in Alaskan tundra and boreal forest. *Journal of Geophysical Research: Earth Surface*, 120(2), 363-378. <http://dx.doi.org/10.1002/2014JF003180>
- Johnston, G. H., Ladanyi, B., Morgenstern, N. R., and Penner, E. (1981). Engineering Characteristics of Frozen and Thawing Soils. In G. H. Johnston (ed.), *Permafrost Engineering Design and Construction*. John Wiley & Sons.
- Johnstone, J. F., Hollingsworth, T. N., Chapin, F. S., and Mack, M. C. (2010). Changes in fire regime break the legacy lock on successional trajectories in Alaskan boreal forest. *Global Change Biology*, 16(4), 1281-1295. <https://doi.org/10.1111/j.1365-2486.2009.02051.x>
- Jones, B. M., Grosse, G., Arp, C. D., Miller, E., Liu, L., Hayes, D. J., and Larsen, C. F. (2015). Recent arctic tundra fire initiates widespread thermokarst development. *Scientific Reports*, 5, 15865. <https://doi.org/10.1038/srep15865>
- Jones, M. C., Booth, R. K., Yu, Z., and Ferry, P. (2013). A 2200-year record of permafrost dynamics and carbon cycling in a collapse-scar bog, interior Alaska. *Ecosystems*, 16, 1-19. <https://doi.org/10.1007/s10021-012-9592-5>
- Jones, M. C., Grosse, G., Jones, B. M., and Anthony, K. W. (2012). Peat accumulation in drained thermokarst lake basins in continuous, ice-rich permafrost, northern Seward Peninsula, Alaska. *Journal of Geophysical Research: Biogeosciences*, 117(G2). <https://doi.org/10.1029/2011JG001766>
- Jorgenson, M. T., Harden, J., Kanevskiy, M., O'Donnell, J., Wickland, K., Ewing, S., Manies, K., Zhuang, Q., Shur, Y., Striegl, R., and Koch, J. (2013). Reorganization of vegetation, hydrology and soil carbon after permafrost degradation across heterogeneous boreal landscapes. *Environmental Research Letters*, 8(3), 035017. <http://dx.doi.org/10.1088/1748-9326/8/3/035017>
- Jorgenson, M. T., Racine, C. H., Walters, J. C., and Osterkamp, T. E. (2001). Permafrost degradation and ecological changes associated with a warming climate in central Alaska. *Climatic Change*, 48(4), 551-579. <https://doi.org/10.1023/A:1005667424292>
- Jorgenson, M. T., Romanovsky, V., Harden, J., Shur, Y., O'Donnell, J., Schuur, E. A. G., Kanevskiy, M., and Marchenko, S. (2010). Resilience and vulnerability of permafrost to climate change. *Canadian Journal of Forest Research*, 40(7), 1219-1236. <https://doi.org/10.1139/X10-060>

- Kasischke, E.S., and Johnstone, J.F. (2005). Variation in postfire organic layer thickness in a black spruce forest complex in interior Alaska and its effects on soil temperature and moisture. *Canadian Journal of Forest Research*, 35, 2164-2177. <https://doi.org/10.1139/X05-159>
- Kasischke, E. S., and Stocks, B. J. (2000). *Fire, Climate Change, and Carbon Cycling in the Boreal Forest*. Springer-Verlag.
- Kasischke, E. S., Verbyla, D. L., Rupp, T. S., McGuire, A. D., Murphy, K. A., Jandt, R., Barnes, J. L., Hoy, E. E., Duffy, P. A., Calef, M., and Turetsky, M. R. (2010). Alaska's changing fire regime - implications for the vulnerability of its boreal forests. *Canadian Journal of Forest Research*, 40(7), 1313-1324. <https://doi.org/10.1139/X10-098>
- Kawahigashi, M., Prokushkin, A., and Sumida, H. (2011). Effect of fire on solute release from organic horizons under larch forest in Central Siberian permafrost terrain. *Geoderma*, 166(1), 171-180. <https://doi.org/10.1016/j.geoderma.2011.07.027>
- Kettridge, N., Humphrey, R. E., Smith, J. E., Lukenback, M. C., Devito, K. J., Petrone, R. M., and Waddington, J. M. (2014). Burned and unburned peat water repellency: implications for peatland evapotranspiration following wildfire. *Journal of Hydrology*, 513, 335-341. <https://doi.org/10.1016/j.jhydrol.2014.03.019>
- Knorre, A. A., Kirdyanov, A. V., Prokushkin, A. S., Krusic, P. J., Büntgen, U. (2019). Tree ring-based reconstruction of the long-term influence of wildfires on permafrost active layer dynamics in Central Siberia. *Science of the Total Environment*, 652, 314-319. <https://doi.org/10.1016/j.scitotenv.2018.10.124>
- Koch, J. C., Kikuchi, C. P., Wickland, K. P., and Schuster, P. (2014). Runoff sources and flow paths in a partially burned, upland boreal catchment underlain by permafrost. *Water Resour Res*, 50(10), 8141-8158. <https://doi.org/10.1002/2014WR015586>
- Kokelj, S. V., and Burn, C. R. (2003). Ground ice and soluble cations in near-surface permafrost, Inuvik, Northwest Territories, Canada. *Permafrost and Periglacial Processes*, 14(3), 275-289. <https://doi.org/10.1002/ppp.458>
- Kong, J., and Yang, J. (2014). Short-and long-term effects of fire on soil properties in a Dahurian larch forest in Great Xing'an Mountains. *Chinese Journal of Ecology*, 33(6), 1445-1450 (in Chinese).
- Kopp, B. J., Lange, J., and Menzel, L. (2017). Effects of wildfire on runoff generating processes in northern Mongolia. *Regional Environmental Change*, 17, 1951-1963. <https://doi.org/10.1007/s10113-016-0962-y>
- Köster, K., Berninger, F., Heinonsalo, J., Lindén, A., Köster, E., Ilvesniemi, H., and Pumpanen, J. (2015). The long-term impact of low-intensity surface fires on litter decomposition and enzyme activities in boreal coniferous forests. *International Journal of Wildland Fire*, 25(2), 213-223. <https://doi.org/10.1071/WF14217>

- Kwok, R. (2018). Arctic sea ice thickness, volume, and multiyear ice coverage: losses and coupled variability (1958–2018). *Environmental Research Letters*, 13(10), 105005. <https://doi.org/10.1088/1748-9326/aae3ec>
- Kwong, Y. J., and Gan, T. Y. (1994). Northward migration of permafrost along the Mackenzie Highway and climatic warming. *Climatic Change*, 26(4), 399-419. <https://doi.org/10.1007/BF01094404>
- Kyle, K. (2015, July 9). Ecology North gets \$130K to study health costs of forest fire smoke. *CBC North*. <https://www.cbc.ca/news/canada/north/ecology-north-gets-130k-to-study-health-costs-of-forest-fire-smoke-1.3144159>
- Lacelle, D., Fontaine, M., Pellerin, A., Kokelj, S. V., and Clark, I. D. (2019). Legacy of Holocene landscape changes on soil biogeochemistry: a perspective from paleo-active layers in northwestern Canada. *Journal of Geophysical Research: Biogeosciences*, 124(9), 2662-2679. <https://doi.org/10.1029/2018JG004916>
- Lantz, T. C., and Kokelj, S. V. (2008). Increasing rates of retrogressive thaw slump activity in the Mackenzie Delta region, NWT, Canada. *Geophysical Research Letters*, 35(6). <https://doi.org/10.1029/2007GL032433>
- Lewis, T. L., Lindberg, M. S., Schmutz, J. A., and Bertram, M. R. (2014). Multi-trophic resilience of boreal lake ecosystems to forest fires. *Ecology*, 95(5), 1253-1263. <https://doi.org/10.1890/13-1170.1>
- Lewkowicz, A. G. (2008). Evaluation of miniature temperature-loggers to monitor snowpack evolution at mountain permafrost sites, northwestern Canada. *Permafrost and Periglacial Processes*, 19(3), 323-331.
- Lewkowicz, A. G., Etzelmüller, B., and Smith, S. L. (2011). Characteristics of discontinuous permafrost based on ground temperature measurements and electrical resistivity tomography, southern Yukon, Canada. *Permafrost and Periglacial Processes*, 22(4), 320-342. <https://doi.org/10.1002/ppp.703>
- Lewkowicz, A. G., and Harris, C. (2005). Morphology and geotechnique of active-layer detachment failures in discontinuous and continuous permafrost. *Geomorphology*, 69(1-4), 275-297. <https://doi.org/10.1016/j.geomorph.2005.01.011>
- Li, X., Jin, H., He, R., Huang, Y., Wang, H., Luo, D., Jin, X., Lü, L., Wang, L., Li, W., Wei, C., Chang, X., Yang, S., and Yu, S. (2019). Effects of forest fires on the permafrost environment in the northern Da Xing'anling (Hinggan) mountains, Northeast China. *Permafrost and Periglacial Processes*, 30(3), 163-177. <https://doi.org/10.1002/ppp.2001>
- Lipovsky, P. S., Coates, J., Lewkowicz, A. G., and Trochim, E. (2005). Active layer detachments following the summer 2004 forest fires near Dawson City, Yukon. In D. S. Edmond, G. D. Bradshaw, L. L. Lewis, and L. H. Weston (Eds.), *Yukon Exploration and Geology* (pp. 175-194). Yukon Geological Survey.

- Liu, H. P., Randerson, J. T., Lindfors, J., Chapin III, F. S. (2005). Changes in the surface energy budget following fire in boreal ecosystems of interior Alaska: an annual perspective. *Journal of Geophysical Research*, 110(D13). <https://doi-org/10.1029/2004JD005158>
- Loke, M.H. (2001). *Electrical imaging surveys for environmental and engineering studies, a practical guide to 2-D and 3-D surveys*. RES2DINV Manual, IRIS Instruments.
- Loke, M. H., Acworth, I., and Dahlin, T. (2003). A comparison of smooth and blocky inversion methods in 2D electrical imaging surveys. *Exploration Geophysics*, 34(3), 182–187. <https://doi.org/10.1071/EG03182>
- Loranty, M. M., Abbott, B. W., Blok, D., Douglas, T. A., Epstein, H. E., Forbes, B. C., Jones, B. M., Kholodov, A. L., Kropp, H., Malhotra, A., Mamet, S. D., Myers-Smith, I. H., Natali, S. M., O'Donnell, J. A., Phoenix, G. K., Rocha, A. V., Sonnentag, O., Tape, K. D., and Walker, A. D. (2018). Reviews and syntheses: Changing ecosystem influences on soil thermal regimes in northern high-latitude permafrost regions. *Biogeosciences*, 15(17), 5287–5313. <https://doi.org/10.5194/bg-15-5287-2018>
- Loranty, M. M., Natali, S. M., Berner, L. T., Goetz, S. J., Holmes, R. M., Davydov, S. P., Zimov, N. S., and Zimov, S. A. (2014). Siberian tundra ecosystem vegetation and carbon stocks four decades after wildfire. *Journal of Geophysical Research: Biogeosciences*, 119(11), 2144–2154. <https://doi.org/10.1002/2014JG002730>
- Mackay, J. R. (1977). Probing for the bottom of the active layer. Geological Survey of Canada, Paper No. 77-1A, 327-328.
- Mackay, J. R. (1995). Active layer changes (1968 to 1993) following the forest-tundra fire near Inuvik, N.W.T., Canada. *Arctic and Alpine Research*, 27(4), 323-336.
- Manasypov, R. M., Shirokova, L. S., and Pokrovsky, O. S. (2017). Experimental modelling of thaw lake water evolution in discontinuous permafrost zone: Role of peat, lichen leaching and ground fire. *Science of the Total Environment*, 580, 245-257. <https://doi.org/10.1016/j.scitotenv.2016.12.067>
- McClymont, A. F., Hayashi, M., Bentley, L. R., and Christensen, B. S. (2013). Geophysical imaging and thermal modeling of subsurface morphology and thaw evolution of discontinuous permafrost. *Journal of Geophysical Research: Earth Surface*, 118(3), 1826–1837. <https://doi.org/10.1002/jgrf.20114>
- McGuire, A. D., Hayes, D. J., Kicklighter, D. W., Manizza, M., Zhuang, Q., Chen, M., Follows, M. J., Gurney, K. R., McClelland, J. W., Melillo, J. M., and Peterson, B. J. (2010). An analysis of the carbon balance of the Arctic Basin from 1997 to 2006. *Tellus B: Chemical and Physical Meteorology*, 62(5), 455-474. <https://doi.org/10.1111/j.1600-0889.2010.00497.x>
- McGuire, A. D., Macdonald, R. W., Schuur, E. A. G., Harden, J. W., Kuhry, P., Hayes, D. J., Christensen, T. R., and Heimann, M. (2010). The carbon budget of the northern

- cryosphere region. *Current Opinion in Environmental Sustainability*, 2(4), 231-236. <https://doi.org/10.1016/j.cosust.2010.05.003>
- Mekonnen, Z. A., Riley, W. J., Randerson, J. T., Grant, R. F., and Rogers, B. M. (2019). Expansion of high-latitude deciduous forests driven by interactions between climate warming and fire. *Nature Plants*, 5, 952-958. <https://doi.org/10.1038/s41477-019-0495-8>
- Michaelides, R. J., Schaefer, K., Zebker, H. A., Parsekian, A., Liu, L., Chen, J., Natali, S., Ludwig, S., and Schaefer, S. R. (2019). Inference of the impact of wildfire on permafrost and active layer thickness in a discontinuous permafrost region using the remotely sensed active layer thickness (ReSALT) algorithm. *Environmental Research Letters*, 14(3), 035007. <https://doi.org/10.1088/1748-9326/aaf932>
- Miller, C. R., Routh, P. S., Brosten, T. R., McNamara, J. P. (2008). Application of time-lapse ERT imaging to watershed characterization. *Geophysics*, 73(3), G7–G17. <https://doi.org/10.1190/1.2907156>
- Minsley, B. J., Pastick, N. J., Wylie, B. K., Brown, D. R. N., and Kass, M. A. (2016). Evidence for nonuniform permafrost degradation after fire in boreal landscapes. *Journal of Geophysical Research: Earth Surface*, 121(2), 320-335. <https://doi.org/10.1002/2015JF003781>
- Morse, P. D., Burn, C. R., and Kokelj, S. V. (2012). Influence of snow on near-surface ground temperatures in upland and alluvial environments of the outer Mackenzie Delta, Northwest Territories. *Canadian Journal of Earth Sciences*, 49(8), 895–913. <https://doi.org/10.1139/e2012-012>
- Myers-Smith, I. H., and Hik, D. S. (2018). Climate warming as a driver of tundra shrubline advance. *Journal of Ecology*, 106(2), 547-560. <https://doi.org/10.1111/1365-2745.12817>
- Natural Resources Canada. (2020). *Precise Point Positioning*. Canadian Geodetic Survey. <https://webapp.geod.nrcan.gc.ca/geod/tools-outils/ppp.php?locale=en>
- Nelson, F. E., Anisimov, O. A., and Shiklomanov, N. I. (2002). Climate change and hazard zonation in the Circum-Arctic permafrost regions. *Natural Hazards*, 26, 203–225. <https://doi.org/10.1023/A:1015612918401>
- Neumann, R. B., Moorberg, C. J., Lundquist, J. D., Turner, J. C., Waldrop, M. P., McFarland, J. W., Euskirchen, E. S., Edgar, C. W., Turetsky, M. R. (2019). Warming effects of spring rainfall increase methane emissions from thawing permafrost. *Geophysical Research Letters*, 46(3), 1393-1401. <https://doi.org/10.1029/2018GL081274>
- Northern Climate Exchange. (2013). *Burwash Landing and Destruction Bay landscape hazards: geological mapping for climate change adaptation planning*. Yukon Research Centre, Yukon College. https://www.yukonu.ca/sites/default/files/inline-files/Hazards_BurwashDBay_finalApr29_web.pdf

- Nossov, D. R., Jorgenson, M. T., Kielland, K., and Kanevskiy, M. Z. (2013). Edaphic and microclimatic controls over permafrost response to fire in interior Alaska. *Environmental Research Letters*, 8(3), 035013. <https://doi.org/10.1088/1748-9326/8/3/035013>
- Obu, J., Westermann, S., Kääb, A., and Bartsch, A. (2018). *Ground Temperature Map, 2000-2016, Northern Hemisphere Permafrost*. PANGAEA, Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, Bremerhaven, Germany. <https://doi.org/10.1594/PANGAEA.888600>.
- O'Donnell, J. A., Harden, J. W., McGuire, A. D., Kanevskiys, M. Z., Jorgenson, M. T., and Xu, X. (2011a). The effect of fire and permafrost interactions on soil carbon accumulation in an upland black spruce ecosystem of interior Alaska: implications for post-thaw carbon loss. *Global Change Biology*, 17(3), 1461-1474. <https://doi.org/10.1111/j.1365-2486.2010.02358.x>
- O'Donnell, J. A., Harden, J. W., McGuire, A. D., and Romanovsky, V. E. (2011b). Exploring the sensitivity of soil carbon dynamics to climate change, fire disturbance and permafrost thaw in a black spruce ecosystem. *Biogeosciences*, 8, 1367-1382. <https://doi.org/10.5194/bg-8-1367-2011>
- O'Donnell, J. A., Jorgenson, M. T., Harden, J. W., McGuire, A. D., Kanevskiy, M. Z., and Wickland, K. P. (2012). The effects of permafrost thaw on soil hydrologic, thermal, and carbon dynamics in an Alaskan Peatland. *Ecosystems*, 15, 213-229. <https://doi.org/10.1007/s10021-011-9504-0>
- Oke, T. R. (1987). *Boundary Layer Climates*. 2nd edition. Methuen, London.
- Oldenborger, G. A., Knoll, M. D., Routh, P. S., LaBrecque, D. J. (2007). Time-lapse ERT monitoring of an injection/withdrawal experiment in a shallow unconfined aquifer. *Geophysics*, 72(4), 177–187. <https://doi.org/10.1190/1.2734365>
- Olefeldt, D., Goswami, S., Grosse, G., Hayes, D., Hugelius, G., Kuhry, P., McGuire, A. D., Romanovsky, V. E., Sannel, A. B. K., Schuur, E. A. G., Turetsky, M. R. (2016). Circumpolar distribution and carbon storage of thermokarst landscapes. *Nature Communications*, 7, 13043. <https://doi.org/10.1038/ncomms13043>
- O'Neill, H. B., Roy-Léveillé, P., Lebedeva, L., and Ling, F. (2020). Recent advances (2010-2019) in the study of taliks. *Permafrost and Periglacial Processes*, 31(3), 346-357. <https://doi.org/10.1002/ppp.2050>
- O'Neill, H. B., Wolfe, S. A., and Duchesne, C. (2019). New ground ice maps for Canada using a paleogeographic modelling approach. *Cryosphere*, 13, 753–773. <https://doi.org/10.5194/tc-13-753-2019>
- O'Neill, H. B., Wolfe, S. A., and Duchesne, C. (2020). *Ground ice map of Canada v.1*. Geological Survey of Canada, Open File 8713. In press.

- Owens, P. N., Giles, T. R., Petticrew, E. L., Leggat, M. S., Moore, R. D., and Eaton, B. C. (2013). Muted responses of streamflow and suspended sediment flux in a wildfire-affected watershed. *Geomorphology*, 202, 128–139. <https://doi.org/10.1016/j.geomorph.2013.01.001>
- Pastick, N. J., Jorgenson, M. T., Wylie, B. K., Minsley, B. J., Ji, L., Walvoord, M. A., Smith, B. D., Abraham, J. D., and Rose, J. R. (2013). Extending airborne electromagnetic surveys for regional active layer and permafrost mapping with remote sensing and ancillary data, Yukon Flats Ecoregion, Central Alaska. *Permafrost and Periglacial Processes*, 24(3), 184–199. <https://doi.org/10.1002/ppp.1775>
- Prowse, T.D. (2011). Lake and river ice in Canada. In: French, H. and O. Slaymaker (Eds.), *Changing Cold Environments*. pp. 163–181. John Wiley and Sons.
- Quinton, W. L., Hayashi, M., and Chasmer, L. E. (2009). Peatland hydrology of discontinuous permafrost in the Northwest Territories: overview and synthesis. *Canadian Water Resources Journal*, 34(4), 311–328. <https://doi.org/10.4296/cwrj3404311>
- Quinton, W. L., Hayashi, M., and Chasmer, L. E. (2011). Permafrost-thaw-induced landcover change in the Canadian subarctic: Implications for water resources. *Hydrological Processes*, 25(1), 152–158. <https://doi.org/10.1002/hyp.7894>
- Reed, D. E., Ewers, B. E., Pendall, E., Frank, J., Kelly, R. (2016). Bark beetle-induced tree mortality alters stand energy budgets due to water budget changes. *Theoretical and Applied Climatology*, 131(1-2), 153–165. <https://doi.org/10.1007/s00704-016-1965-9>
- Riseborough, D. W., and Burgess, M. M. (1996). Measurement interval and the accurate assessment of ground temperature trends. *Permafrost and Periglacial Processes*, 7(4), 321–336. [https://doi.org/10.1002/\(SICI\)1099-1530\(199610\)7:4%3C321::AID-PPP233%3E3.0.CO;2-J](https://doi.org/10.1002/(SICI)1099-1530(199610)7:4%3C321::AID-PPP233%3E3.0.CO;2-J)
- Riseborough, D. W., and Smith, M. W. (1998). Exploring the limits of permafrost. In A. G. Lewkowicz, M. Allard (eds.), *Proceedings of Seventh International Conference on Permafrost* (pp. 935–941). Centre d'études Nordiques, Collection Nordicana, 57.
- Robinson, S. D., and Moore, T. R. (2000). The influence of permafrost and fire upon carbon accumulation in high boreal peatlands, northwest territories, Canada. *Arctic, Antarctic, and Alpine Research*, 32(2), 155–166. <https://doi.org/10.1080/15230430.2000.12003351>
- Rocha, A. V., Loranty, M. M., Higuera, P. E., Mack, M. C., Hu, F. S., Jones, B. M., Breen, A. L., Rastetter, E. B., Goetz, S. J., and Shaver, G. R. (2012). The footprint of Alaskan tundra fires during the past half-century: Implications for surface properties and radiative forcing. *Environmental Research Letters*, 7(4), 044039. <https://doi.org/10.1088/1748-9326/7/4/044039>

- Rocha, A. V., and Shaver, G. R. (2011). Postfire energy exchange in arctic tundra: the importance and climatic implications of burn severity. *Global Change Biology*, 17(9), 2831–2841. <https://doi.org/10.1111/j.1365-2486.2011.02441.x>
- Romanovsky, V. E., Smith, S. L., and Christiansen, H. H. (2010). Permafrost thermal state in the polar Northern Hemisphere during the international polar year 2007–2009: a synthesis. *Permafrost and Periglacial Processes*, 21(2), 106–116. <https://doi.org/10.1002/ppp.689>
- Rouse, W. R. (1976). Microclimatic changes accompanying burning in subarctic lichen woodland. *Arctic and Alpine Research*, 8(4), 357–376.
- Schuur, E. A. G., Bockheim, J., Canadell, J. G., Euskirchen, E., Field, C. B., Goryachkin, S. V., Hagemann, S., Kuhry, P., Lafleur, P. M., Lee, H., Mazhitova, G., Nelson, F. E., Rinke, A., Romanovsky, V. E., Shiklomanov, N., Tarnocai, C., Venevsky, S., Vogel, J. G., and Zimov, S. A. (2008). Vulnerability of permafrost carbon to climate change: implications for the global carbon cycle. *BioScience*, 58(8), 701–714. <https://doi.org/10.1641/B580807>
- Schuster, P. F., Striegl, R. G., Aiken, G. R., Krabbenhoft, D. P., Dewild, J. F., Butler, K., Kamark, B., and Dornblaser, M. (2011). Mercury export from the Yukon River Basin and potential response to a changing climate. *Environmental Science & Technology*, 45(21), 9262–9267. <https://doi.org/10.1021/es202068b>
- Seibert, J., McDonnell, J. J., and Woodsmith, R. D. (2010). Effects of wildfire on catchment runoff response: A modelling approach to detect changes in snow-dominated forested catchments. *Hydrology Research*, 41(5), 378–390. <https://doi.org/10.2166/nh.2010.036>
- Serreze, M. C., and Francis, J. A. (2006). The Arctic amplification debate. *Climatic Change*, 76(3–4), 241–264. <https://doi.org/10.1007/s10584-005-9017-y>
- Serreze, M. C., and Stroeve, J. (2015). Arctic sea ice trends, variability and implications for seasonal ice forecasting. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 373(2045), 20140159. <https://dx.doi.org/10.1098%2Fsta.2014.0159>
- Sharp, M., Burgess, D. O., Cawkwell, F., Copland, L., Davis, J. A., Dowdeswell, E. K., Dowdeswell, J. A., Gardner, A. S., Mair, D., Wang, L., and Williamson, S. N. (2014). Remote sensing of recent glacier changes in the Canadian Arctic. In G. J. Leonard, A. Kääb, J. S. Kargel, B. H. Raup, M. P. Bishop (eds.), *Global Land Ice Measurements from Space* (pp. 205–228). Springer Berlin Heidelberg.
- Shur, Y. L., and Jorgenson, M. T. (2007). Patterns of permafrost formation and degradation in relation to climate and ecosystems. *Permafrost and Periglacial Processes*, 18(1), 7–19. <https://doi.org/10.1002/ppp.582>
- Seguin, M.-K. and Frydecki, J. (1994). Semi-quantitative geophysical investigation of permafrost in northern Québec. *Journal of Applied Geophysics*, 32(1):73–84. [https://doi.org/10.1016/0926-9851\(94\)90010-8](https://doi.org/10.1016/0926-9851(94)90010-8)

- Sjöberg, Y., Marklund, P., Pettersson, R., and Lyon, S.W. (2015). Geophysical mapping of palsa peatland permafrost. *The Cryosphere*, 9:465-478. <https://doi.org/10.5194/tc-9-465-2015>
- Smith, S. L., and Burgess, M. (2004). *Sensitivity of permafrost to climate warming in Canada*. Bulletin of the Geological Survey of Canada (No. 579), Natural Resources Canada.
- Smith, S. L., Burgess, M. M., Riseborough, D., and Nixon, F. M. (2005). Recent trends from Canadian permafrost thermal monitoring network sites. *Permafrost and Periglacial Processes*, 16(1), 19-30. <https://doi.org/10.1002/ppp.511>
- Smith, S. L., Duchesne, C., and Lewkowicz, A. G. (2019). Tracking changes in permafrost thermal state in northern Canada. In J.-P. Bilodeau, D.F. Nadeau, D. Fortier, and D. Conciatori (eds.), *Cold Regions Engineering 2019: Proceedings of the 18th International Conference on Cold Regions Engineering and 8th Canadian Permafrost Conference* (pp. 64-72). American Society of Civil Engineers, Reston, Virginia. <https://doi.org/10.1061/9780784482599.077>
- Smith, M. W., and Riseborough, D. W. (2002). Climate and the limits of permafrost: a zonal approach. *Permafrost and Periglacial Processes*, 13(1), 1–15. <https://doi.org/10.1002/ppp.410>
- Smith, S. L., Riseborough, D. W., and Bonnaventure, P. P. (2015). Eighteen-year record of forest fire effects on ground thermal regimes and permafrost in the central Mackenzie Valley, NWT, Canada. *Permafrost and Periglacial Processes*, 26(4), 289-303. <https://doi.org/10.1002/ppp.1849>
- Smith, S. L., Romanovsky, V. E., Lewkowicz, A. G., Burn, C. R., Allard, M., Clow, G. D., Yoshikawa, K., and Throop, J. (2010). Thermal state of permafrost in North America—a contribution to the International Polar Year. *Permafrost and Periglacial Processes*, 21(2), 117–135. <http://dx.doi.org/10.1002/ppp.690>
- Stocks, B.J., Fosberg, M.A., Lynham, T.J., Mearns, L., Wotton, B.M., Yang, Q., Jin, J.Z., Lawrence, K., Hartley, G.R., Mason, J.A. and McKenney, D.W. (1998). Climate change and forest fire potential in Russian and Canadian boreal forests. *Climatic Change*, 38, 1-13. <https://doi.org/10.1023/A:1005306001055>
- Sun, H., Santalahti, M., Pumpanen, J., Köster, K., Berninger, F., Raffaello, T., Jumpponen, A., Asiegbu, F. O., and Heinonsalo, J. (2015). Fungal community shifts in structure and function across a boreal forest fire chronosequence. *Applied and Environmental Microbiology*, 81, 7869-7880. <https://doi.org/10.1128/AEM.02063-15>
- Swanson, D. K. (1996). Susceptibility of permafrost soils to deep thaw after forest fires in interior Alaska, U.S.A., and some ecologic implications. *Arctic and Alpine Research*, 28(2), 217-227.
- Tank, S. E., Olefeldt, D., Quinton, W. L., Spence, C., Dion, N., Ackley, C., Burd, K., Hutchins, R., and Mengistu, S. (2018). Fire in the Arctic: The effect of wildfire across diverse

- aquatic ecosystems of the Northwest Territories. In *Aqhaliat* (pp. 31-38), Polar Knowledge Canada. <https://www.canada.ca/en/polar-knowledge/aqhaliat.html>
- Tank, S. E., Vonk, J. E., Walvoord, M. A., McClelland, J. W., Laurion, I., and Abbott, B. W. (2020). Landscape matters: Predicting the biogeochemical effects of permafrost thaw on aquatic networks with a state factor approach. *Permafrost and Periglacial Processes*, 31(3), 358-370. <https://doi.org/10.1002/ppp.2057>
- Tarnocai, C., Canadell, J. G., Schuur, E. A. G., Kuhry, P., Mazhitova, G., and Zimov, S. (2009). Soil organic carbon pools in the northern circumpolar permafrost region. *Global Biogeochemical Cycles*, 23(2). <https://doi.org/10.1029/2008GB003327>
- Taş, N., Prestat, E., McFarland, J. W., Wickland, K. P., Knight, R., Berhe, A. A., Jorgenson, T., Waldrop, M. P., and Jansson, J. K. (2014). Impact of fire on active layer and permafrost microbial communities and metagenomes in an upland Alaskan boreal forest. *The ISME Journal*, 8(9), 1904-1919. <https://doi.org/10.1038/ismej.2014.36>
- Throop, J., Lewkowicz, A.G., and Smith, S.L. (2012). Climate and ground temperature relations at sites across the continuous and discontinuous permafrost zones, northern Canada. *Canadian Journal of Earth Sciences*, 49(8), 865–876. <https://doi.org/10.1139/e11-075>
- Tukey, J. W. (1977). Exploratory data analysis. Addison-Wesley Publishing Co.
- Turetsky, M. R., Donahue, W. F., and Benscoter, B. W. (2011). Experimental drying intensifies burning and carbon losses in a northern peatland. *Nature Communications*, 2, 514. <https://doi.org/10.1038/ncomms1523>
- Turetsky, M. R., Mack, M. C., Hollingsworth, T. N., and Harden, J. W. (2010). The role of mosses in ecosystem succession and function in Alaska's boreal forest. *Canadian Journal of Forest Research*, 40(7), 1237-1264. <https://doi.org/10.1139/X10-072>
- Turetsky, M. R., Wieder, R. K., Halsey, L., and Vitt, D. H. (2002). Current disturbance and the diminishing peatland carbon sink. *Geophysical Research Letters*, 29(11), 1-4. <https://doi.org/10.1029/2001GL014000>
- Van Cleve, K., and Viereck, L.A. (1983). A comparison of successional sequences following fire on permafrost-dominated and permafrost-free sites in interior Alaska. In *Proceedings of the 4th International Conference on Permafrost* (pp. 1286-1291). Fairbanks, Alaska, National Academy Press.
- van Everdingen, R. (2005). Multi-language glossary of permafrost and related ground-ice terms. National Snow and Ice Data Center, Boulder, Colorado, USA.
- Veraverbeke, S., Rogers, B. M., Goulden, M. L., Jandt, R. R., Miller, C. E., Wiggins, E. B., and Randerson, J. T. (2017). Lightning as a major driver of recent large fire years in North American boreal forests. *Nature Climate Change*, 7, 529-534. <https://doi.org/10.1038/nclimate3329>

- Verma, S., and Jayakumar, S. (2012). Impact of forest fire on physical, chemical and biological properties of soil: A review. *Proceedings of the International Academy of Ecology and Environmental Sciences*, 2(3), 168-176.
- Verret, M. (2017). *Distribution, morphology and carbon stock of earth hummocks in the Chuck Creek Trail Valley, northern British Columbia, Canada* [M.Sc. thesis]. University of Ottawa.
- Viereck, L. A. (1982). Effects of fire and firelines on active layer thickness and soil temperatures in interior Alaska. In *Proceedings of the 4th Canadian Permafrost Conference, The Roger J. E. Brown Memorial Volume* (pp. 123-134). National Research Council of Canada.
- Viereck, L. A., Werdin-Pfisterer, N. R., Adams, P. C., and Yoshikawa, K. (2009). Effect of wildfire and fireline construction on the annual depth of thaw in a black spruce permafrost forest in interior Alaska: a 36-year record of recovery. In D. L. Kane and K. M. Hinkel (Eds.), *Proceedings of the 9th International Conference on Permafrost*, Institute of Northern Engineering, University of Alaska Fairbanks, Volume 2, 1845-1850.
- Vincent, L. A., Zhang, X., Brown, R. D., Feng, Y., Mekis, E., Milewska, E. J., Wan, H., Wang, X. L. (2015). Observed trends in Canada's climate and influence of low-frequency variability modes. *Journal of Climate*, 28(6), 4545-4560. <https://doi.org/10.1175/JCLI-D-14-00697.1>
- Walker, X. J., Mack, M. C., and Johnstone, J. F. (2015). Stable carbon isotope analysis reveals widespread drought stress in boreal black spruce forests. *Global Change Biology*, 21(8), 3102-3113. <https://doi.org/10.1111/gcb.12893>
- Walker, X. J., Baltzer, J. L., Cumming, S. G., Day, N. J., Johnstone, J. F., Rogers, B. M., Solvik, K., Turetsky, M. R., and Mack, M. C. (2018a). Soil organic layer combustion in boreal black spruce and jack pine stands of the Northwest Territories, Canada. *International Journal of Wildland Fire*, 27(2), 125-134. <https://doi.org/10.1071/WF17095>
- Walker, X. J., Rogers, B. M., Baltzer, J. L., Cumming, S. R., Day, N. J., Goetz, S., Johnstone, J. F., Schuur, E. A. G., Turetsky, M. R., and Mack, M. C. (2018b). Cross-scale controls on carbon emissions from boreal mega-fires of the Northwest Territories, Canada. *Global Change Biology*, 24(9), 4251-4265. <https://doi.org/10.1111/gcb.14287>
- Walker, X. J., Baltzer, J. L., Cumming, S. G., Day, N. J., Goetz, S., Johnstone, J. F., Potter, S., Rogers, B. M., Schuur, E. A. G., Turetsky, M. R., and Mack, M. C. (2019). Increasing wildfires threaten historic carbon sink of boreal forest soils. *Nature*, 572, 520-523. <https://doi.org/10.1038/s41586-019-1474-y>
- Walz, J., Knoblauch, C., Böhme, L., and Pfeiffer, E.-M. (2017). Regulation of soil organic matter decomposition in permafrost-affected Siberian tundra soils – Impact of oxygen

- availability, freezing and thawing, temperature, and labile organic matter. *Soil Biology and Biochemistry*, 110, 34-43. <https://doi.org/10.1016/j.soilbio.2017.03.001>
- Wang, X., Thompson, D. K., Marshall, G. A., Tymstra, C., Carr, R., and Flannigan, M. D. (2015). Increasing frequency of extreme fire weather in Canada with climate change. *Climatic Change*, 130, 573-586. <https://doi.org/10.1007/s10584-015-1375-5>
- Way, R. G. and Lewkowicz, A. G. (2015). Investigations of discontinuous permafrost in coastal Labrador with DC electrical resistivity tomography. In *Proceedings of the 68th Canadian Geotechnical Conference & 7th Canadian Permafrost Conference*. Canadian Geotechnical Society.
- Way, R. G., Lewkowicz, A. G., and Zhang, Y. (2018). Characteristics and fate of isolated permafrost patches in coastal Labrador, Canada. *The Cryosphere*, 12, 2667-2688. <https://doi.org/10.5194/tc-12-2667-2018>
- Williams, P. J., and Smith, M. W. (Eds.). (1989). The Ground Thermal Regime. In *The Frozen Earth: Fundamentals of Geocryology* (pp. 83-121). Cambridge University Press.
- Wolfe, S. A., Duchesne, C., Kokelj, S. V., Stevens, C. W., Gaanderse, A., Houben, A. J., and D'Onofrio, R. E. (2011). *Report on 2010–2011 permafrost investigations in the Yellowknife Area, Northwest Territories*. Geological Survey of Canada.
- Wolfe, S. A., Morse, P. D., Hoeve, T. E., Sladen, W. E., Kokelj, S. V., and Arenson, L. U. (2015). Disequilibrium permafrost conditions on NWT Highway 3. In *Proceedings of the 68th Canadian Geotechnical Conference & 7th Canadian Permafrost Conference*. Canadian Geotechnical Society.
- Woo, M. K. (1990). Permafrost hydrology. In T. D. Prowse, C. Simon, L. Ommanney (Eds.), *Northern Hydrology, Canadian Perspectives, Proceedings of the Northern Hydrology Symposium* (pp. 63-76). National Hydrology Research Institute.
- Wotton, B. M., Flannigan, M. D., and Marshall, G. A. (2017). Potential climate change impacts on fire intensity and key wildfire suppression thresholds in Canada. *Environmental Research Letters*, 12(9), 095003. <https://doi.org/10.1088/1748-9326/aa7e6e>
- Wotton, B. M., Nock, C. A., and Flannigan, M. D. (2010). Forest fire occurrence and climate change in Canada. *International Journal of Wildland Fire*, 19(3), 253-271. <https://doi.org/10.1071/WF09002>
- Yi, S., McGuire, A. D., Kasischke, E., Harden, J., Manies, K., Mack, M., and Turetsky, M. (2010). A dynamic organic soil biogeochemical model for simulating the effects of wildfire on soil environmental conditions and carbon dynamics of black spruce forests. *Journal of Geophysical Research: Biogeosciences*, 115(G4). <https://doi.org/10.1029/2010JG001302>

- Yoshikawa, K., Bolton, W. R., Romanovsky, V. E., Fukuda, M., and Hinzman, L. D. (2002). Impacts of wildfire on the permafrost in the boreal forests of Interior Alaska. *Journal of Geophysical Research: Atmospheres*, 107(D1), FFR 4-1-FFR 4-14. <https://doi.org/10.1029/2001JD000438>
- Yoshikawa, K., Leuschen, C., Ikeda, A., Harada, K., Gogineni, P., Hoekstra, P., Hinzman, L., Sawada, Y., and Matsuoka, N. (2006). Comparison of geophysical investigations for detection of massive ground ice (pingo ice). *Journal of Geophysical Research: Planets*, 111(E06). <https://doi.org/10.1029/2005JE002573>
- Zhang, Y., Chen, W. J., and Riseborough, D. W. (2008). Transient projections of permafrost distribution in Canada during the 21st century under scenarios of climate change. *Global and Planetary Change*, 60(3-4), 443-456. <https://doi.org/10.1016/j.gloplacha.2007.05.003>
- Zhang, Y., Wolfe, S. A., Morse, P. D., Olthof, I., and Fraser, R. H. (2015). Spatiotemporal impacts of wildfire and climate warming on permafrost across a subarctic region, Canada. *Journal of Geophysical Research: Earth Surface*, 120(11), 2338-2356. <https://doi.org/10.1002/2015JF003679>
- Zhou, X., Sun, H., Pumpanen, J., Sietiö, O.-M., Heinonsalo, J., Köster, K., and Berninger, F. (2019). *Biogeochemistry*, 142, 1-17. <https://doi.org/10.1007/s10533-018-0510-6>
- Zoltai, S. C. (1993). Cyclic development of permafrost in the peatlands of northwestern Alberta, Canada. *Arctic and Alpine Research*, 25(3), 240-246.

APPENDICES

APPENDIX A: SUPPLEMENTAL INFORMATION FOR CHAPTER 2

Table A1: Data for post-fire changes to the thermal regime of permafrost from recent (2010-2019) literature, including the residual organic layer thickness (OLT) and the time since the sites burned. Sites have been grouped into boreal forest and tundra, and organized sequentially by years since fire. Differences between burned and unburned locations for active layer thickness (ALT), ground surface temperature (GST), and ground temperature (GT) at 1 m depth are reported. Blank spaces indicate no data reported. Timing and depth of temperature measurements is given in brackets.

Study	Location	Years Since Fire	OLT (cm)	ALT (cm)	Δ ALT (cm)	Δ GST (°C)	Δ GT (°C)
Li et al., 2019	Boreal forest, valley, northeast China	1	20	130	20	- 2.8 (mean annual)	- 0.2 (mean annual T @ 100 cm)
Li et al., 2019	Boreal forest, valley, northeast China	1	45	130	20	- 1.6 (mean annual)	- 0.8 (mean annual T @ 100 cm)
Nossov et al., 2013	Boreal forest, silty upland, Alaska	1	7	125	66	+ 3.3 (mean annual)	+ 3 (mean annual T @ 100 cm)
Knorre et al., 2019	Boreal forest, central Siberia	1-2		150			
Gibson et al., 2018	Peat plateaus, boreal forest, western Canada	1-5	200- > 300	85	35		
Douglas et al., 2016	Boreal forest, lowland, Alaska	2	Thin	83.3- 169.7			
Nossov et al., 2013	Boreal forest, sandy lowland, Alaska	2-6	< 2	196	87	+ 3.7 (mean annual)	+ 1.5 (mean annual T @ 100 cm)
Smith et al., 2015	Boreal forest, hillslope, western Canada	2-11	0-15				+ 2 (mean T @ 100 cm)

Jiang et al., 2015	Boreal forest, upland, Alaska	4	14			- 0.3-1.6 (mean summer)	+ 4.9-5.4 (mean summer T @ 20 cm)
Smith et al., 2015	Boreal forest, hillslope, western Canada	4	0-15			+ 7 (mean winter)	
Smith et al., 2015	Boreal forest, hillslope, western Canada	5	0-15			+ 1 (mean annual)	
Kopp et al., 2014	Boreal forest, slope, Mongolia	5	3-5			+ 3.9 (mean annual)	+ 4.3 (mean annual T @ 30 cm)
Fisher et al., 2016	Boreal forest, western Canada	6	~ 30 – 60	100-140	~ 60-100		
Zhang et al., 2015	Boreal forest, western Canada	6	10-30	97-121	59-68		
Nossov et al., 2013	Boreal forest, rocky upland, Alaska	6	4-9	> 100	> 51	+ 2 (mean annual)	+ 2.1 (mean annual T @ 100 cm)
Li et al., 2019	Boreal forest, valley, northeast China	7	25	350	300	+ 3.7 (mean annual)	+ 0.1 (mean annual T @ 100 cm)
Li et al., 2019	Boreal forest, valley, northeast China	7	40	100	50	+ 1.8 (mean annual)	+ 0.64 (mean annual T @ 100 cm)
Bai et al., 2013	Boreal forest, mild burn, Mongolia	7				+ 1.8 (mean summer)	+ 0.64 (mean summer T @ 20 cm)
Bai et al., 2013	Boreal forest, severe burn, Mongolia	7				+ 5.38 (mean summer)	+ 2.3 (mean summer T @ 20 cm)

Gibson et al., 2018	Peat plateaus, boreal forest, western Canada	10-20	200->300	60-70	10-20		+ 6-7 (max T @ 40 cm)
Douglas et al., 2016	Boreal forest, lowland, Alaska	11	14	114->250			
Li et al., 2019	Boreal forest, valley, northeast China	15	30	200	50	+ 1.7 (mean annual)	+ 2.6 (mean annual T @ 100 cm)
Jafarov et al., 2013	Boreal forest, Alaska	17-21	30-80	400	340		
Douglas et al., 2016	Boreal forest, lowland, Alaska	24	176	56.3-66.4			
Li et al., 2019	Boreal forest, valley, northeast China	30	30	150	0	- 3.2 (mean annual)	+ 1.2 (mean annual T @ 100 cm)
Douglas et al., 2016	Boreal forest, lowland, Alaska	37	7	> 250			
Jiang et al., 2015	Upland boreal forest, Alaska	40	15			+ 0.8-1.5 (mean summer)	+ 3-3.4 (mean summer T @ 20 cm)
Douglas et al., 2016	Boreal forest, lowland, Alaska	82	89	53.3-63.3			
Jiang et al., 2015	Tundra moderate burn, Alaska	1	15.4				+ 1.3-2 (mean summer T @ 20 cm)
Jiang et al., 2015	Tundra severe burn, Alaska	1	11.6			+ 1 (mean summer)	+ 2.4-3.2 (mean summer T @ 20 cm)

Rocha et al., 2011	Tundra, moderate burn, Alaska	1-3	16.2	50	6	+ 1.5-2.3 (mean summer)	
Rocha et al., 2011	Tundra, severe burn, Alaska	1-3	12.5	60	12	+ 3.7-4 (mean summer)	
Jones et al., 2015	Tundra, centre of ice wedge polygon, Alaska	2	0-5				+ 0.2 (mean September T @ 100 cm)
Jones et al., 2015	Tundra, centre of ice wedge polygon, Alaska	6	0-5				+ 0.6 (mean September T @ 100 cm)
Rocha et al., 2012	Tundra, Alaska	10			6		
Barrett et al., 2012	Tundra, Alaska	17		43	6		
Loranty et al., 2014	Tundra, northeastern Siberia	41			No significant difference		

Notes: +, Higher; -, Lower. T = temperature.

Appendix A2: Supplementary references

1. Treat CC, Wisser D, Marchenko S, Forlking S. Modelling the effects of climate change and disturbance on permafrost stability in northern organic soils. *Mires Peat*. 2013;12(2):1-17.
2. Portier J, Gauthier S, Leduc A, Arseneault D, Bergeron Y. Fire regime along latitudinal gradients of continuous to discontinuous coniferous boreal forests in eastern Canada. *Forests*. 2016;7(10).
3. Beck PSA, Goetz SJ, Mack MC, et al., The impacts and implications of an intensifying fire regime on Alaskan boreal forest composition and albedo. *Glob Change Biol*. 2011;17:2853–2866.
4. Wotton BM, Nock CA, Flannigan MD. Forest fire occurrence and climate change in Canada. *Int J Wildland Fire*. 2010;19:253-271.
5. Jain P, Wang X, Flannigan MD. Trend analysis of fire season length and extreme fire weather in North America between 1979 and 2015. *Int J Wildland Fire*. 2018;26(12):1009-1020.
6. Rocha AV, Shaver GR. Postfire energy exchange in arctic tundra: the importance and climatic implications of burn severity. *Glob Change Biol*. 2011;17:2831–2841.
7. Hu FS, Higuera PE, Walsh JE, Chapman WL, Duffy PA, Brubaker LB. Tundra burning in Alaska: Linkages to climatic change and sea ice retreat. *J Geophys Res-Bioge*. 2010;115(G4).
8. Brown DRN, Jorgenson MT, Kielland K, Verbyla DL, Prakash A, Koch JC. Landscape effects of wildfire on permafrost distribution in interior Alaska derived from remote sensing. *Remote Sensing*. 2016;8:654.
9. Knorre AA, Kirdyanov AV, Prokushkin AS, Krusic PJ, Büntgen U. Tree ring-based reconstruction of the long-term influence of wildfires on permafrost active layer dynamics in Central Siberia. *Sci Total Environ*. 2019;652:314-319.
10. Zhuang Q, McGuire AD, Melillo JM, et al., Carbon cycling in extratropical terrestrial ecosystems of the Northern Hemisphere during the 20th century: A modelling analysis of the influences of soil thermal dynamics. *Tellus B*. 2011;55(3):751-776.
11. O'Neill KP, Kasischke ES, Richter DD. Seasonal and decadal patterns of soil carbon uptake and emission along an age sequence of burned black spruce stands in interior Alaska. *J Geophys Res-Atmos*. 2003;108(D1):1-15.
12. Shi L, Dech JP, Liu H, Zhao P, Bayin D, Zhou M. Post-fire vegetation recovery at forest sites is affected by permafrost degradation in the Da Xing'an Mountains of northern China. *J Veg Sci*. 2019;00:1-10.
13. Mann DH, Rupp TS, Olson MA, Duffy PA. Is Alaska's boreal forest now crossing a major ecological threshold? *Arct Antarct Alp Res*. 2012;44:319-331.
14. Johnstone JF, Chapin FS, Hollingsworth TN, Mack MC, Romanovsky V, Turetsky M. Fire, climate change, and forest resilience in interior Alaska. *Can J Forest Res*. 2010a;40(7):1302-1312.

15. Dyrness CT. *Control of depth to permafrost and soil temperature by the forest floor in black spruce/feathermoss communities*. Portland, OR: USDA, Forest Service, Pacific Northwest Forest and Range Experiment Station, Research Note, PNW-396. 1982;19 pp.
16. Kasischke ES, Stocks BJ. *Fire, Climate Change, and Carbon Cycling in the Boreal Forest*. Springer-Verlag, New York; 2000.
17. Liljedahl A, Hinzman L, Busey R, Yoshikawa K. Physical short-term changes after a tussock tundra fire, Seward Peninsula, Alaska. *J Geophys Res-Earth*. 2007;112:F02S07.
18. MacKay JR. Disturbances to the tundra and forest tundra environment of the western Arctic. *Can Geotech J*. 1970;7:420-432.
19. Viereck LA. Effects of fire and firelines on active layer thickness and soil temperatures in interior Alaska. In: French HM, ed. *Proceedings of the 4th Canadian Permafrost Conference, The Roger J. E. Brown Memorial Volume*. Ottawa, ON: National Research Council of Canada; 1982:123-135.
20. Jiang Y, Zhuang Q, O'Donnell JA. Modelling thermal dynamics of active layer soils and near-surface permafrost using a fully coupled water and heat transport model. *J Geophys Res*. 2012;117:1-15.
21. Walker XJ, Rogers BM, Baltzer JL, Cumming SR, Day NJ, Goetz S, Johnstone JF, Schuur EAG, Turetsky MR, Mack MC. Cross-scale controls on carbon emissions from boreal mega-fires of the Northwest Territories, Canada. *Glob Change Biol*. 2018b;24:4251-4265.
22. Whitman E, Parisien M-A, Thompson DK, Hall RJ, Skakun RS, Flannigan MD. Variability and drives of burn severity in the northwestern Canadian boreal forest. *Ecosphere*. 2018;9(2):e02128.
23. Thompson DK, Parisien M-A, Morin J, Millard K, Larsen CPS, Simpson BN. Fuel accumulation in a high-frequency boreal wildfire regime: from wetland to upland. *Can J For Res*. 2017;47(7):957-964.
24. Sofronov MA, Volokitina AV. Wildfire ecology in continuous permafrost zone. In: Osawa A, Zyryanova OA, Matsuura Y, Kajimoto T, Wein RW, eds. *Permafrost Ecosystems - Siberian Larch Forests*. Dordrecht, Heidelberg, London, New York: Springer Science+Business Media; 2010:59-82.
25. Kopp BJ, Minderlein S, Menzel L. Soil moisture dynamics in a mountainous headwater area in the discontinuous permafrost zone of northern Mongolia. *Arct Antarct Alp Res*. 2014;46(2):459-70.
26. Barrett K, Rocha AV, Van de Weg MJ, Shaver G. Vegetation shifts observed in arctic tundra 17 years after fire. *Remote Sens Lett*. 2012;3(8):729-736.
27. Smith MW, Riseborough DW. Climate and the limits of permafrost: a zonal analysis. *Permafrost Periglac*. 2002;13:1-15.
28. Harpold AA, Biederman JA, Condon K, Merino M, Korgaonkar Y, Nan T, Sloat LL, Ross M, Brooks PD. Changes in snow accumulation and ablation following the Las Conchas forest fire, New Mexico, USA. *Ecohydrol*. 2014;7:440-452.

29. Winkler R. Changes in snow accumulation and ablation after a fire in south-central British Columbia. *Watershed Management Bulletin*. 2011;14(2):1-7.
30. Semenova O, Lebedeva L, Volkova N, et al., Detecting immediate wildfire impact on runoff in a poorly-gauged mountainous permafrost basin. *Hydrol Sci J*. 2015;60(7-8):1225-1241.
31. Jones BM, Baughman CA, Romanovsky VE, et al., Presence of rapidly degrading permafrost plateaus in south-central Alaska. *Cryosphere*. 2016;10(6):2673-2692.
32. Lara MJ, Genet H, McGuire AD, et al., Thermokarst rates intensify due to climate change and forest fragmentation in an Alaskan boreal forest lowland. *Glob Change Biol*. 2016;22:816–829.
33. Liu L, Jafarov EE, Schaefer KM, et al., InSAR detects increase in surface subsidence caused by an Arctic tundra fire. *Geophys Res Lett*. 2014;41(11):3906-3913.
34. Zoltai SC. Cyclic development of permafrost in the peatlands of northwestern Alberta, Canada. *Arctic Alpine Res*. 1993;25(3):240-246.
35. Jorgenson MT, Roth JE, Smith MD, et al., *An ecological land survey for Fort Greely, Alaska*. Hanover, NH: U.S. Army Cold Regions Research and Engineering Laboratory, TR-01-04; 2001.
36. Larouche JR, Abbott BW, Bowden WB, Jones JB. The role of watershed characteristics, permafrost thaw, and wildfire on dissolved organic carbon biodegradability and water chemistry in Arctic headwater streams. *Biogeosciences*. 2015;12(14):4221-4233.
37. Hanna AJ, McNeill D, Tchekhovski A, Fridel T, Babkirk C. 1998. The effects of the 1994 and 1995 forest fires on the slopes of the Norman Wells pipeline. In: Lewkowicz AG, Allard M, eds. *Proceedings of the Seventh International Conference on Permafrost*. Yellowknife, Canada: Collection Nordicana, Centre d'etudes Nordiques, Quebec City; 1998:421-426.
38. Hughes OL, Veillette JJ, Pilon J, Hanley PT, van Everington RO. *Terrain evaluation with respect to pipeline construction, Mackenzie Transportation Corridor, Central Part, lat 64° to 68°N*. Ottawa, Canada: Environmental-Social Committee, Northern Pipelines, Task Force on Northern Oil Development No. 73-37; 1973.
39. Lewkowicz AG, Bonnaventure PP. Equivalent elevation: a new method to incorporate variable lapse rates into mountain permafrost modelling. *Permafrost Periglac*. 2011;22:153–162.
40. Héon J, Arseneault D, Parisien M-A. Resistance of the boreal forest to high burn rates. *P Nat A Sci*. 2014;111(38) :13888-13893.
41. Mack MC, Bret-Harte, MS, Hollingsworth TN, et al., Carbon loss from an unprecedented Arctic tundra wildfire. *Nature*. 2011;475:489-492.
42. Bernhardt EL, Hollingsworth TN, Chapin FS. Fire severity mediates climate-driven shifts in understory community composition of black spruce stands of interior Alaska. *J Veg Sci*. 2011;22(1):32-44.

43. Onodera SI, Van Stan JT. Effect of forest fires on hydrology and biogeochemistry of watersheds. In: Levina DF, Carlyle-Moses D, Tanakas T, eds. *Forest Hydrology and Biogeochemistry*. Dordrecht, Netherlands: Springer; 2011:599-621.
44. Olefeldt D, Devito KJ, Turetsky MR. Sources and fate of terrestrial dissolved organic carbon in lakes of a Boreal Plains region recently affected by wildfire. *Biogeoosci.* 2013;10(10):6247-6265.
45. Stoof CR, Wesseling JG, Ritsema CJ. Effects of fire and ash on soil water retention. *Geoderma*. 2010;159(3-4):276-285.
46. Quinton WL, Hayashi M, Chasmer LE. Permafrost-thaw-induced land-cover change in the Canadian subarctic: implications for water resources. *Hydrol Process*. 2011;25(1):152-158.
47. Sherwood JH, Kettridge N, Thompson DK, Morris PJ, Silins U, Waddington JM. Effect of drainage and wildfire on peat hydrophysical properties. *Hydrol Process*. 2013;27(13):1866-1874.
48. Parham LM, Prokushkin AS, Pokrovsky OS, et al., Permafrost and fire as regulators of stream chemistry in watersheds of the Central Siberian Plateau. *Biogeochemistry*. 2013;116:55-68.
49. Allison SD, Lu Y, Weihe C, et al., Microbial abundance and composition influence litter decomposition response to environmental change. *Ecology*. 2013;94:714-725.
50. Dooley SR, Treseder KK. The effect of fire on microbial biomass: a meta-analysis of field studies. *Biogeochem*. 2012;109(1-3):49-61.
51. Waring BG, Averill C, Hawkes CV. Differences in fungal and bacterial physiology alter soil carbon and nitrogen cycling: Insights from meta-analysis and theoretical models. *Ecol Lett*. 2013;16:887-894.
52. Xu X, Thornton PE, Post WM. A global analysis of soil microbial biomass carbon, nitrogen and phosphorus in terrestrial ecosystems. *Glob Ecol Biogeogr*. 2013;22:737-749.
53. Singh BK, Bardgett RD, Smith P, Reay DS. Microorganisms and climate change: Terrestrial feedbacks and mitigation options. *Nat Rev Microbiol*. 2010;8:779-790.
54. Jassey VEJ, Recuga MK, Zielińska M, Slowińska S, et al., Tipping point in plant-fungal interactions under severe drought causes abrupt rise in peatland ecosystem respiration. *Glob Change Biol*. 2018;24(3):972-986.
55. Silins U, Bladon KD, Kelly EN, Esch E, Spence JR, Stone M, et al., Five-year legacy of wildfire and salvage logging impacts on nutrient runoff and aquatic plant, invertebrate, and fish productivity. *Ecohydrology*. 2014;7:1508-1523.
56. Keuper F, van Bodegom PM, Dorrepaal E, Weedon JT, van Hal J, van Logtestijn RSP, Aerts R. A frozen feast: Thawing permafrost increases plant-available nitrogen in subarctic peatlands. *Glob Change Biol*. 2012;18(6):1998-2007.

57. Smith HG, Sheridan GJ, Lane PNJ, Nyman P, Haydon S. Wildfire effects on water quality in forest catchments: A review with implications for water supply. *J Hydrol.* 2011;396:170–192.
58. McEachern P, Prepas EE, Gibson JJ, Dinsmore WP. Forest fire induced impacts on phosphorus, nitrogen, and chlorophyll *a* concentrations in boreal subarctic lakes of northern Alberta. *Can J Fish Aquat Sci.* 2000;57:73-81.
59. Harden JW, Koven CD, Ping C-L, Hugelius G, McGuire AD, et al., Field information links permafrost carbon to physical vulnerabilities of thawing. *Geophys Res Lett.* 2012;39(15).
60. Hugelius G, Routh J, Kuhry P, Crill P. Mapping the degree of decomposition and thaw remobilization potential of soil organic matter in discontinuous permafrost terrain. *J Geophys Res Biogeosci.* 2012;117(G2).
61. Köster K, Berninger F, Lindén A, Köster E, Pumpanen J. Recovery in fungal biomass is related to decrease in soil organic matter turnover time in a boreal fire chronosequence. *Geoderma.* 2014;235-236:74-82.
62. Myers-Smith IH, Elmendorf SC, Beck PSA, Wilmking M, et al., Climate sensitivity of shrub growth across the tundra biome. *Nat Clim Change.* 2015;5(9):887.
63. Myers-Smith IH, Forbes BC, Wilmking M, Hallinger M, et al., Shrub expansion in tundra ecosystems: dynamics, impacts and research priorities. *Environ Res Lett.* 2011;6(4):045509.

APPENDIX B: SUPPLEMENTAL INFORMATION FOR CHAPTER 3

Table B1: Frozen ground conditions and other observations for all sites where Brown provided this information in 1962. Site numbers are from Brown (1964). Latitude and longitude are in decimal degrees (datum WGS84). For sites where permafrost was probable in 2017-2018, predicted frost table depths are noted with a * (see Table B2). Observed median frost tables for 2017-2018 are presented, with adjusted values in parentheses. PF = permafrost, OL = organic layer. Dashes (-) indicate no data.

Site	Lat	Long	PF Present 1962	PF Present 2017/18	Mean FT Depth 1962 (cm)	Median FT Depth 2017/18 (cm)	PF Thickness 1962 (m)	PF Thickness 2017/18 (m)	Relief	Soil Type	Canopy Surface 1962	Canopy Surface 2017/18	OL Thickness 1962 (cm)	OL Thickness 2017/18 (cm)	Burn Year
1	60.82131	-115.79008	No	Not visited	-	-	-	-	High	-	Jack pine, poplar, spruce	Built-up area	-	-	-
2	60.81249	-115.79851	No	Not visited	-	-	-	-	High	-	Jack pine, poplar, spruce	Built-up area	-	-	-
3	60.75907	-115.86160	Yes	Yes	84	44 (49)	-	8	Low	Clay, silt, with some stones	Spruce; sphagnum	Spruce; moss, lab tea, willow, birch.	91	> 85	1971
4	60.71378	-115.88960	No	No	-	-	-	-	High	Sand and gravel	Poplar, jack pine, spruce	Poplar, spruce, tamarack; Moss, willow, lab tea	-	36	-
5	60.71464	-115.8943	Yes	Probable	84	146*	-	?	Slope	Sand	Spruce, tamarack; moss	Spruce, tamarack; moss, lab tea, horsetails, birch.	-	25	1971
6	60.70697	-115.90380	No	No	-	-	-	-	High	Gravel	Jack pine, poplar, spruce	Poplar, spruce, willow	-	30	1981
7	60.67968	-115.95303	No	Yes	-	60	-	-	River or stream valley	-	-	Spruce, tamarack, willow; grass, labrador tea, moss	-	> 60	-
9	60.6603	-116.01446	Yes	No	54	0	-	0	Low	Sand	Wet, hummocky moss with lichen	Spruce, tamarack (small); Wet, sedge, willow, birch	51	-	-
11	60.62586	-116.06358	No	No	-	-	-	-	Low	Sand	Spruce, tamarack; hummocky sphagnum	Spruce, tamarack; hummocky sphagnum, lichen, lab tea	61	15	-
14	60.60585	-116.09553	No	No	-	-	-	-	High	Gravel	Poplar, jack pine, spruce	Poplar, jack pine, spruce	-	15	-

18	60.57663	-116.11031	Yes	No	113	0	-	0	Low	Sand	Spruce and tamarack; hummocky sphagnum	Spruce and tamarack; hummocky moss, grass, lab tea, lichen, birch	60-152	20	-
19	60.56790	-116.11639	No	No	-	-	-	-	Low	Sand	Spruce; Hummocky moss, lichen, sphagnum	Spruce, tamarack; hummocky moss, horse tails, lab tea	46	38	-
20	60.55503	-116.15326	No	No	-	-	-	-	Sloping	Sand	Spruce, jack pine, tamarack; Moss, lichen, sedge	Spruce, jack pine, tamarack; Moss, lichen, sedge	15	25	-
22	60.52057	-116.22886	No	No	-	-	-	-	High	Sand, silt, clay, some stones	Spruce, jack pine, poplar; Moss, grass, sedge, lichen	Poplar, spruce, jack pine	-	16	-
23	60.49422	-116.29705	Yes	Probable	60	143*	-	?	Low	Organic, sand, silt, clay	Spruce and tamarack; Peat plateau with hummocky moss, lichen, and labrador tea	Spruce and tamarack; moss, horsetails, willow	56	24	-
24	60.49135	-116.30062	No	No	-	-	-	-	Low	Silt	Spruce, tamarack, willow; Wet moss	Spruce, tamarack, willow; Wet moss	-	65	-
25	60.48296	-116.31800	No	No	-	-	-	-	Low	Silt, sand, some stones	Spruce, tamarack, willow; Moss, sedge, willow	Tamarack, willow, spruce; sedge, moss, wet	61	64	-
26	60.46990	-116.32596	Yes	Probable	60	138*	0.2	?	Sloping	Sand and gravel	Spruce; Hummocky moss, lichen, grass	Spruce; hummocky moss, grass, cranberry, rose	61	55	-
28	60.44796	-116.34259	No	No	-	-	-	-	High	Clay, silt, some stones	Jack pine, poplar; Forest litter	Jack pine, poplar; forest litter	-	6	-
31	60.41035	-116.36451	Yes	Yes	84	45 (51)	-	-	Low	Sand, silt	Spruce; hummocky sphagnum and lab tea	Spruce; hummocky moss, lab tea	91	> 60	-

32	60.33794	-116.46609	Yes	No	69	0	>2.4	0	Low	Scattered stones	Spruce; hummocky sphagnum with lichen and lab tea	Spruce, jack pine, poplar; moss, willow	236	70	-
33	60.27453	-116.56459	Yes	Probable	60	129*	-	?	Low	Wet gravel, clay	Spruce, willow; Sphagnum, lichen, and lab tea	Spruce, willow, birch, tamarack; moss, lab tea, sedge, wet	> 61	70	-
34	60.27686	-116.56364	Yes	No	150	0	-	0	Low	Silt, sand, clay	Spruce; hummocky sphagnum, lichen, lab tea	Spruce, willow, birch, tamarack; moss, lab tea, sedge, wet	30	36	-
35	60.27436	-116.56295	Yes	No	61	0	-	0	Low	Silt, sand, clay	Spruce; hummocky sphagnum, lichen, lab tea	Spruce, hummocky moss, lichen, lab tea	30	36	-
36	60.20906	-116.64973	No	No	-	-	-	-	High	Sand and gravel	Poplar, birch, jack pine	Poplar, birch, spruce, jack pine	-	14	-
37	60.17967	-116.68008	No	No	-	-	-	-	High	Silt, sand	Jack pine, spruce, poplar	Poplar, spruce	-	19	-
38	60.10029	-116.76576	No	No	-	-	-	-	High	Organic sand, silt, clay	Poplar, birch, jack pine, spruce; forest litter	Poplar, birch, spruce, tamarack, willow; moss, lab tea	-	27	-
39	60.03948	-116.87433	Yes	Yes	60	61 (71)	> 1	5 to 8	Low	Clay, silt	Spruce; sphagnum	Spruce, tamarack, jack pine; hummocky sphagnum, lichen, lab tea	137	120	-
41	59.9962	-116.98960	Yes	Yes	60	46 (54)	> 0.6	-	Low	Clay, silt, sand	Spruce; Hummocky sphagnum, lichen, and labrador tea	Spruce; hummocky sphagnum and moss, lichen, lab tea, cranberry	91	80	-
43	59.94423	-117.02242	Yes	Yes	-	51 (54)		5 to 8	Low	-	Spruce; sphagnum	Spruce, tamarack; sphagnum, moss, lichen, lab tea	-	> 48	-
44	59.95136	-117.02227	Yes	Not visited	76	-	-	-	Sloping	-	Spruce; hummocky sphagnum	?	-	?	-

45	59.95052	-117.02260	Yes	No	60	0	-	0	Sloping	Clay	Spruce; hummocky sphagnum	Spruce, jack pine, poplar; hummocky moss,	> 61	-	-
46	59.94542	-117.02686	Yes	Yes	60	48 (56)		-	Low	Silt, clay	Spruce; hummocky sphagnum	Spruce; hummocky sphagnum, lichen, lab tea	91	> 45	-
47	59.94048	-117.02804	No	Not visited	-	-	-	-	Low	Clay, silt	Spruce; sphagnum wet	-	61	-	-
48	59.93747	-117.02764	No	No	-	-	-	-	Low	Clay, silt	Spruce; sphagnum, wet	Spruce, tamarack, willow; Moss	61	60	-
49	59.92448	-117.03551	Yes	Yes	66	49 (58)	> 1	-	Low	Clay, silt	Spruce; Peat plateau with hummocky sphagnum and labrador tea	Spruce; hummocky sphagnum, lichen, lab tea	152	80	-
50	59.92441	-117.03552	No	Yes	-	56 (60)	-	-	Low	Clay, silt	Spruce; moss, sedge, wet	Spruce; sedge, wet	30	40	-
51	59.90224	-117.04190	No	No	-	-	-	-	High	Silt, clay	Spruce, tamarack; moss, lichen	Spruce, tamarack, poplar; moss, lichen, grass, rose	76	33	-
52	59.90078	-117.04226	No	No	-	-	-	-	High	Silt, clay	Spruce, poplar	Spruce, poplar; moss, forest litter	30	18	-
53	59.89345	-117.04266	Yes	Yes	61	79 (95)	> 0.3	-	Low	Clay, silt	Spruce; Hummocky sphagnum, lichen, and labrador tea	Spruce; sphagnum, lichen, lab tea, horsetail	91	30	-
54	59.87871	-117.04252	Yes	Yes	61	52 (63)		5	Low	Clay	Spruce; hummocky sphagnum	Spruce; hummocky sphagnum and moss, lichen, lab tea	> 60	40	-
55	59.85770	-117.03937	Yes	No	244	0	> 2.4	0	River or stream valley	Sand	-	Poplar, willow; grass, roses	-	15	-
56	59.82221	-117.04533	Yes	Yes	76	52 (62)	> 1	-	Low	Clay, silt	Spruce; hummocky sphagnum	Spruce, tamarack; hummocky sphagnum, lichen, lab tea	183	132	-

57	59.81100	-117.05457	Yes	Yes	61	47 (53)	-	-	Low	-	Spruce; hummocky sphagnum	Spruce, tamarack, birch; hummocky sphagnum and moss, lichen, lab tea, grass	> 61	> 65	2012
58	59.79554	-117.06028	Yes	Yes	61	48 (55)	-	-	Low	Clay, silt	Spruce; hummocky sphagnum	Spruce, tamarack, birch; hummocky sphagnum, lichen, lab tea	61	> 60	2012
59	59.77014	-117.07363	No	No	-	-	-	-	High	Clay	Poplar, jack pine, spruce	Poplar, spruce; forest litter	-	1	2012
60	59.75018	-117.08062	No	Probable	-	144*	-	?	Sloping	Sand, clay, silt	Spruce; hummocky moss, lichen	Spruce, willow; hummocky moss, horse tails, lab tea, wet	23	46	-
61	59.73900	-117.08811	No	No	-	-	-	-	High	Sand and gravel	Poplar, jack pine	Poplar, spruce	-	15	-
62	59.67410	-117.15090	Yes	Probable	61	110*	0.6	?	Low	Silt, sand	Spruce; Hummocky sphagnum with labrador tea	Spruce; moss and sphagnum, lichen, lab tea, horsetail, grass	91	46	-
63	59.61908	-117.17265	Yes	Yes	61	64 (67)	> 0.6	5	Low	Organic, silt, sand	Spruce; Hummocky sphagnum, lichen, lab tea	Spruce, hummocky sphagnum, moss, lab tea, lichen	102	55	-
64	59.61836	-117.17267	Yes	No	76	0	-	0	Low	Sand	Spruce; sphagnum	Spruce; hummocky moss, lab tea, lichen.	76	30	-
64A	59.61836	-117.17267	Yes	Yes	76		-	-	Low	Sand	Spruce; sphagnum	Spruce; hummocky moss and sphagnum, lab tea, lichen	76	> 50	-
65	59.58442	-117.19044	Yes	Yes	61	65 (73)	-	-	Low	Silt, sand	Spruce; hummocky sphagnum	Spruce, tamarack; hummocky sphagnum, moss, lab tea	61	30	-
66	59.57549	-117.19389	Yes	No	366	0	> 1.2	0	River or stream valley	Gravel	Poplar, spruce, jackpine	Spruce, poplar, willow; roses, forest litter	-	7	-

67	59.54106	-117.18603	No	Yes	-	-	-	-	Low	Silt, clay	Spruce; hummocky sphagnum, lab tea, wet	Spruce; hummocky moss, lab tea, lichen	76	40	-
68	59.52972	-117.18673	Yes	Yes	61	55 (63)	> 0.6	-	Low	Silt, clay	Spruce; peat plateau with hummocky sphagnum and labrador tea	Spruce; hummocky sphagnum and moss, lichen, lab tea	112	> 50	2000/2012
69	59.44915	117.23153	Yes	Yes	51	110 (126)	-	-	Low	Silt, clay	Spruce; Hummocky sphagnum and labrador tea	Spruce; hummocky sphagnum and moss, lab tea	76	140	2007
70	59.40496	-117.28500	Yes	No	-	0	-	0	River or stream valley	Gravel and clay	-	Poplar, willow; moss, fire weed	-	10	-
71	59.38885	-117.30077	Yes	No	-	0	-	0	Low	Clay, Silt	Spruce; sphagnum, peat plateau	Spruce, birch; hummocky sphagnum and lab tea, peat plateau	-	23	2012
72	59.37919	-117.30682	Yes	Yes	61	102 (116)	-	5	Low	-	Spruce; sphagnum	Spruce, birch; hummocky sphagnum, lab tea	> 61	> 50	2012
73	59.29908	-117.37759	Yes	Yes	61	110 (126)	> 3.5	-	Low	Organic, silt, clay	Spruce; hummocky sphagnum with lichen and labrador tea	Spruce; hummocky sphagnum, lab tea	282	> 80	2012
74	59.20103	-117.53072	Yes	No	61	0	-	0	Low	Clay, sand, gravel	Sphagnum, lichen, lab tea	Grass	> 61	25	-
75	59.19257	-117.54987	Yes	Yes	-	61 (65)	-	5	Low	Silt, sand, gravel	Spruce; sphagnum	Spruce; sphagnum, moss	-	> 50	-
76	59.15883	-117.64358	Yes	Yes	61	51 (58)	> 0.2	-	Low	Clay, silt	Spruce; hummocky sphagnum with lichen and labrador tea	Spruce; hummocky moss, lichen, lab tea	79	70	-

78	59.10526	-117.67385	Yes	Yes	-	67 (72)	-	-	Low	Silt, clay, sand	Spruce, tamarack; sphagnum, lab tea	Spruce, tamarack; hummocky sphagnum, lab tea	> 61	80	-
80	59.00904	-117.66958	Yes	No	61	0	> 0.46	0	Sloping	Gravel	Spruce, poplar; hummocky sphagnum, lichen, lab tea	Poplar; moss. Possibly disturbed	107	0	-
81	58.97108	-117.61462	No	No	-	-	-	-	Low	Gravel	Poplar, jack pine, spruce	Poplar, jack pine, spruce	-	-	-
82	58.92592	-117.54022	No	No	-	-	-	-	Sloping	Clay, silt, gravel	Spruce; sphagnum	Spruce; sphagnum, lichen, lab tea, cranberry, wet	122	> 80	-
83	58.79031	-117.39533	Yes	Probable	61	194*	> 0.5	?	Low	Gravel	Spruce; hummocky sphagnum	Spruce, tamarack; sphagnum, sedge, lab tea, cranberry, cloudberry	114	> 80	-
85	58.00439	-117.19946	Yes	No	76	0	0.69	0	Low	Sand, organics	Spruce; hummocky sphagnum, lichen, lab tea	Spruce; hummocky sphagnum, lichen, lab tea	183	> 80	-
86	57.78964	-117.66701	Yes	No	91	0	-	0	Low	-	Spruce; sphagnum	Spruce; sphagnum, lichen, lab tea	91	-	-
87	57.78851	-117.69053	Yes	Not visited	84	-	-	-	Low	-	Spruce; sphagnum	Farmland	> 84	-	-
88	57.78733	-117.71267	Yes	No	84	0	1.3	0	Low	Organic, silt, clay	Hummocky sphagnum, peat plateau	Poplar	127	2	-

Table B2: Thermal gradient measurements and linear regression analysis for sites where no frost table was identified with probing. Thermal gradient measurements were not taken at sites 32, 34, 45, 64, 80, and 88.

Thermal Gradient							
Site	Deepest Measurement Depth (cm)	Measured Temperature (°C)	n	Equation ^a	R ²	Predicted depth of 0°C	Standard Error (cm)
4	150	4.0	4	$y = -52.96x + 365.55$	0.98	>200 cm	6
5	80	3.7	3	$y = -19.55x + 145.63$	0.92	146 cm	10
6	150	9.0	4	$y = -47.70x + 581.40$	0.99	>200 cm	5
9	80	9.9	3	$y = -13.75x + 210.19$	0.92	>200 cm	10
11	120	7.8	4	$y = -36.64x + 398.21$	0.94	>200 cm	10
14	130	6.7	4	$y = -100.30x + 800.46$	0.97	>200 cm	7
18	80	9.2	3	$y = -35.81x + 408.28$	1	>200 cm	0
19	150	5.5	4	$y = -69.93x + 533.24$	1	>200 cm	1
20	150	6.8	4	$y = -62.21x + 575.34$	0.97	>200 cm	7
22	130	6.6	4	$y = -67.60x + 576.25$	0.99	>200 cm	4
23	80	3.0	3	$y = -20.63x + 142.65$	1	143 cm	2
24	140	4.7	4	$y = -48.70x + 369.60$	1	>200 cm	2
25	150	1.9	4	$y = -53.14x + 252.18$	0.99	>200 cm	4
26	80	4.7	3	$y = -13.93x + 137.72$	0.89	138 cm	12
28	150	5.8	4	$y = -36.97x + 368.37$	0.86	>200 cm	15
33	80	4.1	3	$y = -13.04x + 128.75$	0.95	129 cm	8
35	120	0.7	4	$y = -171.54x + 244.39$	0.99	>200 cm	3
36	150	7.3	4	$y = -44.49x + 478.74$	0.93	>200 cm	11
37	150	5.6	4	$y = -28.30x + 311.34$	0.96	>200 cm	8
38	150	3.2	4	$y = -32.96x + 254.82$	1	>200 cm	1
48	150	4.0	4	$y = -41.63x + 319.38$	0.97	>200 cm	7
51	150	3.008	4	$y = -47.02x + 287.68$	0.95	>200 cm	9

52	150	5.122	4	$y = -48.46x + 394.95$	0.93	>200 cm	10
55	120	6.631	4	$y = -56.89x + 499.30$	0.95	>200 cm	8
59	150	7.896	4	$y = -45.21x + 505.58$	1	>200 cm	2
60	120	0.999	4	$y = -27.41x + 144.36$	0.98	144 cm	6
61	150	5.908	4	$y = -44.63x + 414.02$	0.9	>200 cm	12
62	80	1.944	3	$y = -17.59x + 110.35$	0.96	110 cm	7
66	150	5.988	4	$y = -34.67x + 356.95$	1	>200 cm	2
70	150	5.329	4	$y = -46.21x + 393.71$	0.99	>200 cm	4
71	80	6.669	3	$y = -27.85x + 265.66$	1	>200 cm	0
74	150	10.87	4	$y = -56.38x + 766.48$	0.97	>200 cm	7
81	150	5.294	4	$y = -54.68x + 441.45$	0.99	>200 cm	4
82	150	2.824	4	$y = -27.62x + 226.42$	1	>200 cm	2
83	150	0.797	4	$y = -56.89x + 193.81$	0.98	194 cm	6
85	150	1.571	4	$y = -67.03x + 257.90$	0.99	>200 cm	4
86	150	3.063	4	$y = -21.24x + 215.67$	1	>200 cm	1

^aIn the equation, y is depth (cm) and x is temperature (°C)

APPENDIX C: SUPPLEMENTAL INFORMATION FOR CHAPTER 5

Table C1: Study sites established in the southwest NWT in 2015 and 2016. AT = air temperature, GT = ground temperature, GST = ground surface temperature (5 cm below ground surface), BH = borehole, SS = snow stake. Permafrost conditions are those interpreted from ERT surveys in 2015. Sites are presented from south to north, with the two unburned sites last. The 11 sites where detailed permafrost monitoring was conducted are noted with an asterisk (*).

Site name	Latitude	Longitude	Strata	Description	Instrumentation	Date Installed	Permafrost Conditions
*SS33-9a	60.94149	-117.36552	Taiga Plains South, conifer low density	Mesic-subhygric. Peat plateau, lots of thermokarst. No slope.	4 X 21 ERT array; GT at 4-1 (20, 40, 60, 80 cm) and 3-14 (10, 30, 50, 70 cm); BH at 2-19 installed to 3.1 m; BH in unburned installed to 2.4 m; 4 cores	16/06/2015	15-20 m
SS33-6C	60.9575	-117.3389	Taiga Plains South, conifer medium density	Subhygric, spruce forest, lots of mounding. No slope.	4 X 21 ERT array; AT and SS at 2-17; GST at 1-6 and 2-17; GT at 2-17 (15, 50, 70, 90 cm) and 1-9 in unburned (15, 50, 70, 90cm)	04/06/2015	Seasonal frost
ZF20-34b	61.6075	-117.15356	Taiga Plains Central, late burn, jack pine	Mesic-subxeric. Gradient from pine to spruce to bog. No slope.	4 X 21 ERT array; AT and SS at 3-1; GT at 3-1 (25, 50, 75, 100 cm), and at 2-3 (25, 50, 75, 100 cm) and snow stake in unburned.	02/06/2015	Seasonal frost
ZF20-6b-1	61.68932	-116.95173	Taiga Plains Central, early burn, black spruce	Mesic-subxeric. Severely burned, all trees down. No slope.	4 X 21 ERT array; GST at 3-1	08/06/2015	Possible 10 m (patch); not conclusive
*ZF20-6b-LK	61.69166	-116.95312	Taiga Plains Central, early burn, black spruce	Mesic-subhygric. Wet spruce forest near pond; one end has thick peat plateau and the other is a very wet marsh	4 X 21 ERT array; BH at 3-1 (0.5, 0.75, 1, 1.5, 2, 2.5, 3, 3.2 m); SS at 3-1; GT at 2-1 (25, 50, 75, 100 cm); 1 core	11/06/2015	15 m
ZF20-12b	61.73588	-116.83356	Taiga Plains Central, early burn, jack pine	Mesic. Gradient from pine to black spruce to bog. No slope.	4 X 21 ERT array; AT and SS at 3-1; GT at 3-1 (15, 30, 45, 60 cm); 1 core	14/06/2015	Possible 5 m (small patch); not conclusive
ZF20-9c	61.79576	-116.74025	Taiga Plains Central, early burn, black spruce	Mesic-subhygric. Spruce forest with lots of mounds, and small section of unburned. No slope.	4 X 21 ERT array; GT at 2-1 (20, 40, 60, 80 cm) and 1-1 (20, 40, 60, 80 cm); GST at 1-1; 4 short cores	10/06/2015	5-10 m; not conclusive

*ZF20-4a	61.86268	-116.59698	Taiga Plains Central, early burn, black spruce	Mesic-subxeric. Gradient of burn severity from unburned to severely burned. No slope.	4 X 21 ERT array; AT at 3-1; GT in unburned at 2-12 (20, 40, 60, 80 cm) and 4-11 (20, 40, 60, 80 cm)	02/06/2015	10-20 m
*ZF46-18b	62.34465	-116.48166	Taiga Plains North. New burn, late.	Mesic-subhygric. Pine with small area of black spruce. No slope.	4 X 21 ERT array; GST at 3-1 and 4-1; SS at 3-1; GT at 2-13 (15, 30, 45, 60 cm)	09/06/2015	>5 m
ZF46-35c	62.4497	-116.49603	Taiga Plains North. Re-burn, early.	Mesic. Pine and dwarf black spruce patch. No slope.	4 X 21 ERT array; GT (20, 40, 60, 80 cm) and SS at 3-1	09/06/2015	>5 m; not conclusive
ZF46-1b-1	62.48028	-116.51846	Taiga Plains North. New burn, early.	Mesic-subhygric. Small valley, pine on top of hill, spruce in valley bottom (wet/boggy). Hill slope is 30° and valley bottom (60 m) is flat. Sandy.	4 X 21 ERT array; GST at 3-1	06/06/2015	Possible 5 m (small patch); not conclusive
*ZF46-1b-2	62.48088	-116.52031	Taiga Plains North. New burn, early.	Mesic-subhygric. Small valley, pine on top of hill, spruce in valley bottom (wet/boggy). Transect goes through burned and unburned area. Hill slope is 30° and valley bottom (60m) is flat. Sandy.	2 X 21 ERT array; AT and SS at 2-7; BH 1 in unburned at 1-16 (50, 75, 100, 150 cm), BH 2 at 2- 7 (0.5, 1, 1.5, 1.95 m); GST at 1- 15; 2 cores	06/06/2015	5-10 m
*ZF46-1c	62.48494	-116.51777	Taiga Plains North. New burn, early.	Mesic. Spruce stand. Sandy. No slope.	2 X 21 ERT array; GT (20, 40, 60, 80 cm) and SS at 2-1, and 2-5 in unburned (10, 20, 40, 60 cm)	10/06/2015	5-10 m
*TIB-1	62.5222	-113.382	Taiga Shield	Burned peatland with ice wedges	4 X 21 ERT array; 2 cores	05/06/2016	>25 m
*LL-LITH2	63.13243	-114.00634	Taiga Shield. Conifer, low.	Lithalsa topped with spruce and birch trees	4 X 21 ERT array	18/06/2015	>25 m
*LL-PEAT	63.13341	-114.00413	Taiga Shield. Conifer, low.	Peatland in valley next to lithalsas	4 X 21 ERT array	18/06/2015	>25 m
*HWY3- UB-1A	62.02892	-116.31390	Taiga Plains, Unburned	Mesic-subhygric. Unburned forest. No slope.	4 X 21 ERT array; 1 core; BH to 1.5 m	18/05/2016	>25 m

*UB-1	62.45825	-114.53159	Taiga Shield, Unburned	Unburned peat plateau	4 X 21 ERT array; AT and SS at 3-1; BH (7.5 m); 1 core	07/06/2015	>25 m
-------	----------	------------	---------------------------	-----------------------	---	------------	-------

Table C2: Infilled and collated thermal data for all sensors at the study sites.

Site	Year	MAAT	MAGST Unburn	MAGST Burn	MAGT Burn (as close to 1m)	MAGT Unburn (as close to 1m)	Surface Offset Burn	Thermal Offset Burn	Surface Offset Unburn	Thermal Offset Unburn
SS33-9a	2016	-0.7	4.3	4.3	0.0	0.0	5.0	-4.3	5.0	-4.3
SS33-9a	2017	-1.0	3.2	2.8	0.1	0.0	3.9	-2.7	4.2	-3.2
SS33-9a	2018	-2.3	3.3	4.1	0.7	-0.1	6.1	-3.2	5.5	-3.4
SS33-9a	2019	-2.3	2.4	1.8	0.0		4.1	-1.8	4.7	
ZF20-4a	2016	-1.3		4.3	1.6	0.1	5.5	-2.7		
ZF20-4a	2017	-2.1	2.3	3.9	1.6	0.1	6.0	-2.2	4.4	-2.2
ZF20-4a	2018	-3.1			2.5	0.2				
ZF20-4a	2019	-3.2	2.3	3.1		0.1	6.3		5.5	-2.2
ZF20-6b-LK	2016	-0.9		2.9	-0.2		3.8	-3.2		
ZF20-6b-LK	2017	-1.6		2.6	0.6		4.2	-1.9		
ZF20-6b-LK	2018	-2.6								
ZF20-6b-LK	2019	-2.8		3.9	3.1		6.7	-0.8		
ZF46-1b-2	2016	-2.5			-0.2	0.6				
ZF46-1b-2	2017	-3.3	3.2	1.7	-0.2	2.5	5.0	-1.9	6.5	-0.7
ZF46-1b-2	2018	-4.2			-0.4	2.1				
ZF46-1b-2	2019	-4.5	2.3	1.1	0.0	1.9	5.6	-1.2	6.7	-0.4
ZF46-1c	2016	-2.5	2.3	2.7			5.2		4.7	
ZF46-1c	2017	-3.3	2.5	3.5			6.8		5.8	
ZF46-1c	2018	-4.2	3.0	4.0	3.7	1.4	8.2	-0.3	7.3	-1.7
ZF46-1c	2019	-4.5	2.3	3.2	3.0	1.2	7.7	-0.1	6.7	-1.0
ZF46-18b	2016	-2.2		3.0	0.7		5.2	-2.3		

ZF46-18b	2017	-3.0		2.5	1.0		5.6	-1.5
ZF46-18b	2018	-4.0		1.6				
ZF46-18b	2019	-4.2		1.5			5.8	
TIB-1	2017	-2.7		2.9	-0.2		5.6	-3.1
TIB-1	2018	-3.9		1.3	-0.8		5.1	-1.3
TIB-1	2019	-4.3		1.4	-0.1		5.8	-1.5
LL-PEAT	2016	-3.3		1.8			5.1	
LL-PEAT	2017	-5.7		0.7	-1.1		5.1	-0.5
LL-PEAT	2018	-5.0		-0.4	-2.4		4.6	-2.0
LL-PEAT	2019	-5.6		-0.5	-2.5		5.1	-2.0
LL-LITH2	2016	-3.3		0.2			3.5	
LL-LITH2	2017	-5.7		-1.0	-1.1		4.7	-0.1
LL-LITH2	2018	-5.0		-1.9	-1.9		3.0	0.0
LL-LITH2	2019	-5.6		-2.5	-1.8		3.0	0.8
UB-1	2016	-2.2				-0.8		
UB-1	2017	-2.6	1.7			-0.8	4.3	-2.6
UB-1	2018	-3.9	1.5			-1.0	5.3	-2.5
UB-1	2019	-4.3	2.2			-0.6	6.5	-2.7
HWY3-UB-1A	2017	-2.6	1.3			-0.9	3.9	-2.2
HWY3-UB-1A	2018	-3.5	-1.0			-1.3	2.4	-0.2
HWY3-UB-1A	2019	-3.6	0.1			-0.6	3.7	-0.7

Table C3: Detailed soil characteristics for each site. Samples were collected in 2016, except for some used for gravimetric moisture content (GMC) which were collected in 2017. For GMC, the year has been specified in brackets. Two pits with different stratigraphy were sampled at sites ZF46-1c and ZF46-18b, and the pit number is noted in brackets next to the site name.

Site	Depth (cm)	Mean grain size (μm)	% Gravel	% Sand	% Silt	% Clay	Bulk Density (g/cm^3)	GMC (%)	Organic Content (%)	Excess Ice (%)
SS33-9a	24-39	Organic					0.2			
SS33-9a	39-53	Organic					0.2			
SS33-9a	55-65	Organic					0.2			
SS33-9a	65-78	Organic					0.1			
SS33-9a	78-94	Organic					0.1			
SS33-9a	94-100	Organic					0.1			10
SS33-9a	100-117	Organic					0.1			
SS33-9a	117-129	Organic					0.1			
SS33-9a	145-159	Organic					0.1			
SS33-9a	159-174	Organic					0.1			
SS33-9a	174-189	Organic					0.1			
SS33-9a	189-203	Organic					0.1			
SS33-9a	203-214	Organic					0.1			
SS33-9a	214-225	Organic					0.1			
SS33-9a	225-238	Organic					0.1			
SS33-9a	238-251	Organic					0.1			
SS33-9a	251-263	Organic					0.1			
SS33-9a	263-275	Organic					0.2			
SS33-9a	275-297	55.3		25.6	64	10.4	0.2		21.1	10
SS33-9a	297-310	20.6		7.4	66	26.6	0.8		6.9	9
ZF20-4a	15	Organic						336 (2016)		
ZF20-4a	35-40	205.7	48.9	33.3	16.3	1.5		6.6 (2016) 12.4 (2017)	2.5	
ZF20-4a	40	185.3	27.5	47.1	21.1	4.3			2	
ZF20-6b-LK	25	56		31.1	61	8		17 (2017)	1.7	
ZF20-6b-LK	35-40	52.9		27.2	64.3	8.5		22.7 (2017)	2.2	
ZF20-6b-LK	43-58	54.3		28.9	63.6	7.6		58.8 (2016)	1.3	33
ZF20-6b-LK	50	63.6		40.7	52.4	6.9		30 (2016) 27.6 (2017)	3.8	
ZF20-6b-LK	58-72	65.1		41.3	54.1	4.6		41.4 (2016)	1.2	20
ZF20-6b-LK	72-86	85.6		50.8	45.8	3.4		39.7 (2016)	0.9	20
ZF46-1b-2 (1)	25	460		96	4			12.7 (2017)	0.5	

ZF46-1b-2 (2)	38	Organic						426.8 (2017)		
ZF46-1c	15	Organic						485 (2016)		
ZF46-1c	30	563.6	15.9	78.7	5.4			19.2 (2016) 21.2 (2017)	0.8 (2016) 0.5 (2017)	
ZF46-1c	50			81.1	17.2	1.7				
ZF46-18b (1)	10-15	218.1	8.8	66.5	19.6	5.2		6.9 (2017)		
ZF46-18b (2)		Organic						332 (2016)		
ZF46-18b (2)	35	Organic						482.6 (2017)		
ZF46-18b (2)	75								1	
TIB-1	0-18	Organic						123 (2016)		
TIB-1	18-31	Organic						316 (2016)		
TIB-1	31-64	Organic						466.7 (2016)		
TIB-1	50	Organic						334.3 (2017)		
TIB-1	64-93	Organic						569.1 (2016)		
TIB-1	93-99	Organic						587.5 (2016)		
TIB-1	99-124	Organic						371.5 (2016)		
TIB-1	124-153	Organic						519.6 (2016)		
TIB-1	153-167	Organic						237.1 (2016)		
TIB-1	167-199	Organic						491.9 (2016)		
TIB-1	199-218	Organic						263.5 (2016)		
TIB-1	239-254	Organic						321.5 (2016)		
TIB-1	254-289	Organic						278 (2016)		
TIB-1	289-312	25.5		9.9	57.5	32.6		119.6 (2016)	7.3	42
TIB-1	312-325	58.3		29.5	44.1	26.3		81.8 (2016)	3.6	31
HWY3-UB-1A	0-15	Organic						428.8 (2016)		
HWY3-UB-1A	15-20	Organic						12306.9 (2016)		
HWY3-UB-1A	20-35	Organic						736.4 (2016)		
HWY3-UB-1A	35-47	Organic						313.4 (2016)		
HWY3-UB-1A	47-58	30.5		14.9	63.7	21.4		78.9 (2016)	12.4	
HWY3-UB-1A	58-70	31		15.7	61.9	22.4		59.1 (2016)	2.7	
HWY3-UB-1A	60							379.5 (2017)		
HWY3-UB-1A	70-82	42.7		26.6	59.6	13.8		65.2 (2016)	1.8	
HWY3-UB-1A	82-91	40.3		21.3	65.6	13.1		60.7 (2016)	2	
HWY3-UB-1A	91-103	26.9		11.3	72.2	16.5		46.3 (2016)	1.6	
HWY3-UB-1A	103-115	23.8		8.5	73.5	18		46.3 (2016)	1.7	
HWY3-UB-1A	115-121	32.7		14.4	68.6	16.9		86.7 (2016)	2.2	
HWY3-UB-1A	121-133	31.4		13.3	69.4	17.3		87.2 (2016)	2.2	
HWY3-UB-1A	133-143	33.3		14.2	72.4	13.5		88.8 (2016)	2	

HWY3-UB-1A	143-153			16.5	71.5	11.9			1.4	
UB-1	0-4	Organic						153.2 (2016)		
UB-1	4-9	Organic						654.8 (2016)		
UB-1	9-19	Organic						700 (2016)		
UB-1	25-33	Organic						759.4 (2016)		
UB-1	33-50	Organic						379.2 (2016)		
UB-1	45-50	Organic						537 (2017)		
UB-1	50-60	Organic						650 (2016)		
UB-1	60-70	Organic						980.6 (2016)		
UB-1	70-81	Organic						1066.7 (2016)		
UB-1	81-90	Organic						794.6 (2016)		
UB-1	90-99	Organic						805.4 (2016)		
UB-1	99-110	Organic						695.8 (2016)		
UB-1	110-120	Organic						700 (2016)		
UB-1	120-130	Organic						605.7 (2016)		
UB-1	130-145	Organic						634.7 (2016)		
UB-1	145-161	Organic						574.4 (2016)		
UB-1	161-175	Organic						596.5 (2016)		
UB-1	175-185	Organic						555.2 (2016)		
UB-1	185-195	Organic						387.2 (2016)		
UB-1	218-239	Organic						469.1 (2016)		

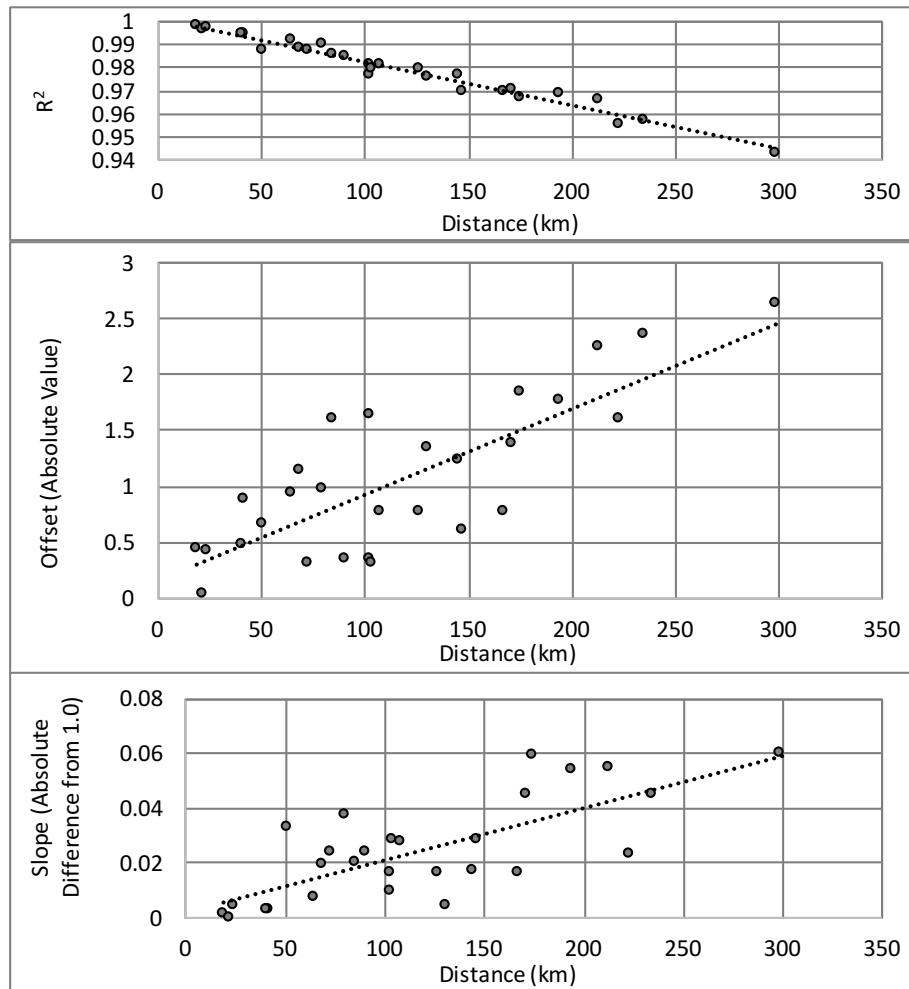


Figure C1: Regression analysis between mean daily air temperature at different sites from June 16, 2015 to September 5, 2019.

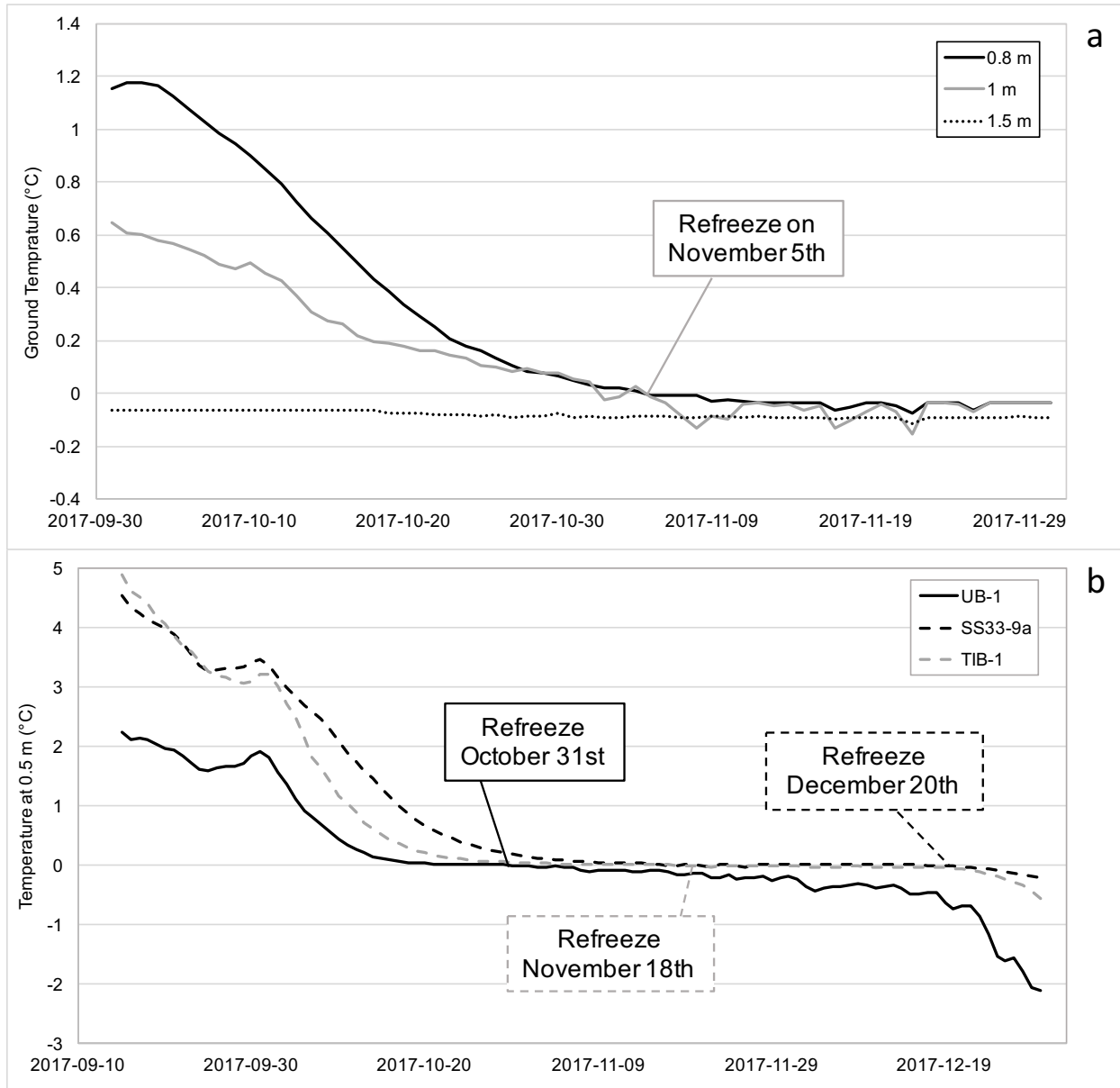


Figure C2: Details on re-freezing conditions within the study area. (a) Ground temperatures at a burned site (SS33-9a) showing the freeze-up period in 2017 for various depths within the soil column. Refreeze for the 1 m sensor occurred on November 5, 2017. (b) The timing of re-freezing at 0.5 m depth for burned (dashed) and unburned (solid) peatlands in the study area, showing that burned sites refreeze weeks later than the unburned site.

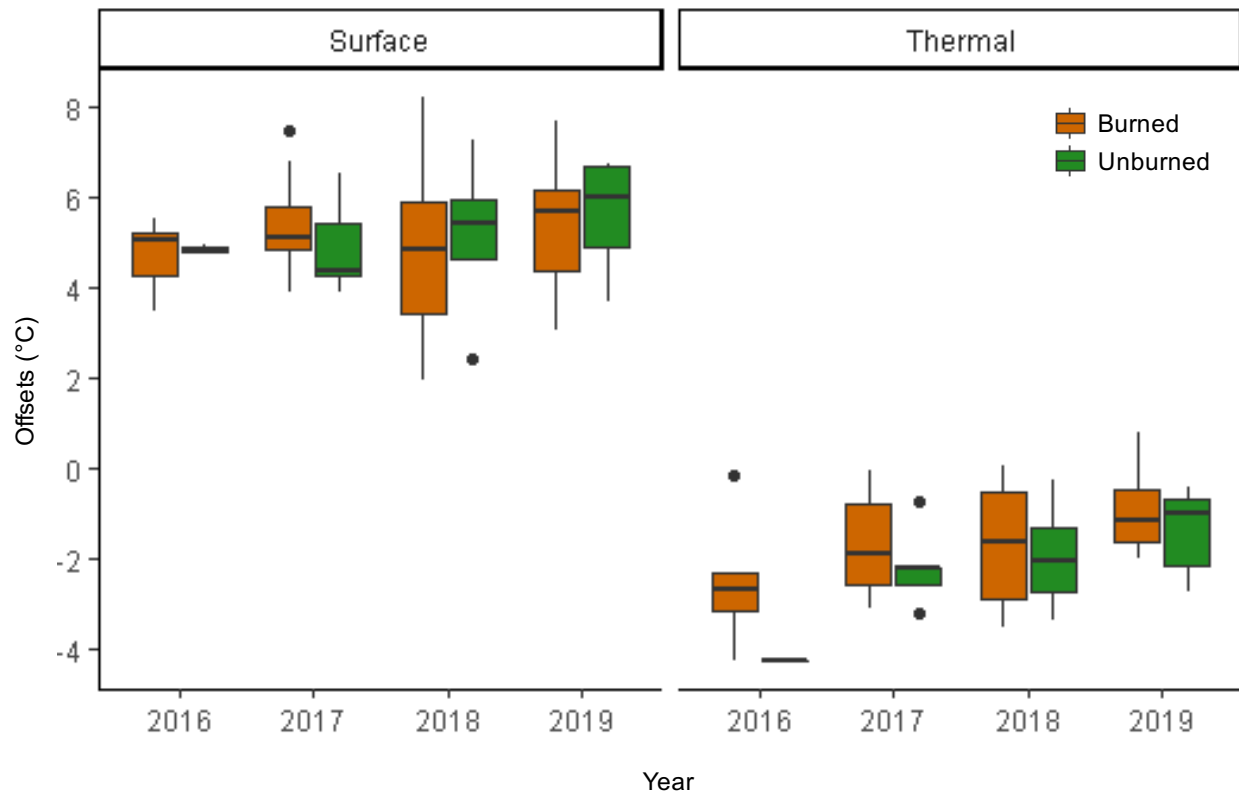


Figure C3: Surface and thermal offsets for all burned and unburned sites, as well as unburned sections of burned sites, between 2016 and 2019. For the surface offset, burned $n = 6-11$ and unburned $n = 2-6$. For the thermal offset, burned $n = 5-10$ and unburned $n = 1-5$. Boxes extend from lower hinge (first quartile) to the upper hinge (third quartile) with the solid line as the median. Whiskers extend to the largest or smallest value no further than $1.5 \times \text{IQR}$ from the hinge (where IQR is the inter-quartile range; Tukey, 1977). Dots beyond the whiskers are outliers.

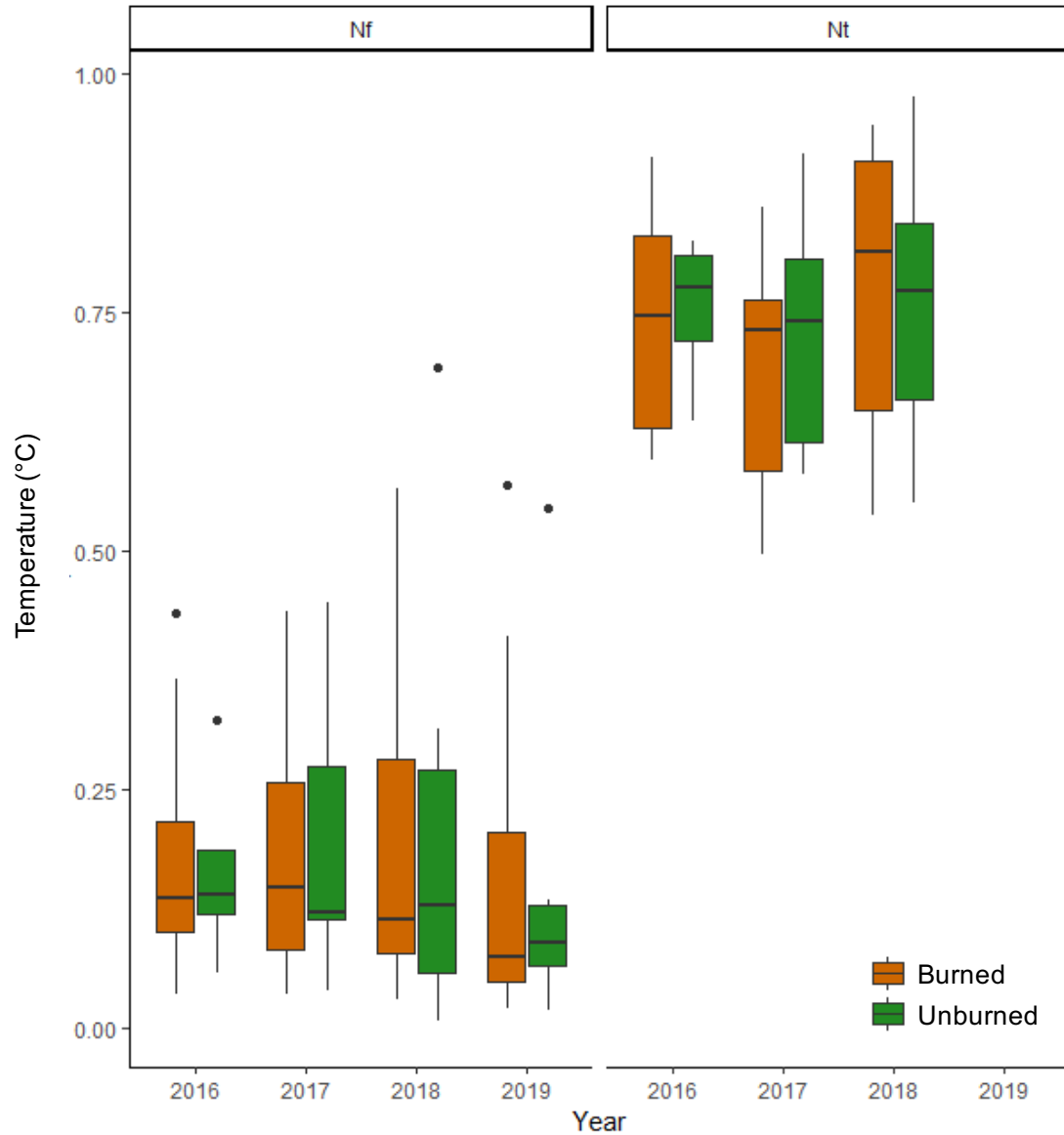


Figure C4: Freezing (N_f) and thawing (N_t) n-factors for all burned and unburned sites, as well as unburned sections of burned sites, in the southern Northwest Territories between 2016 and 2019. No N_t was calculated for 2019 as the measured record stopped before the end of the thawing season. For N_f , burned $n = 8-9$ and unburned $n = 4-6$. For N_t , burned $n = 7-9$ and unburned $n = 4-6$. Boxes extend from lower hinge (first quartile) to the upper hinge (third quartile) with the solid line as the median. Whiskers extend to the largest or smallest value no further than $1.5 * IQR$ from the hinge (where IQR is the inter-quartile range; Tukey, 1977). Dots beyond the whiskers are outliers.

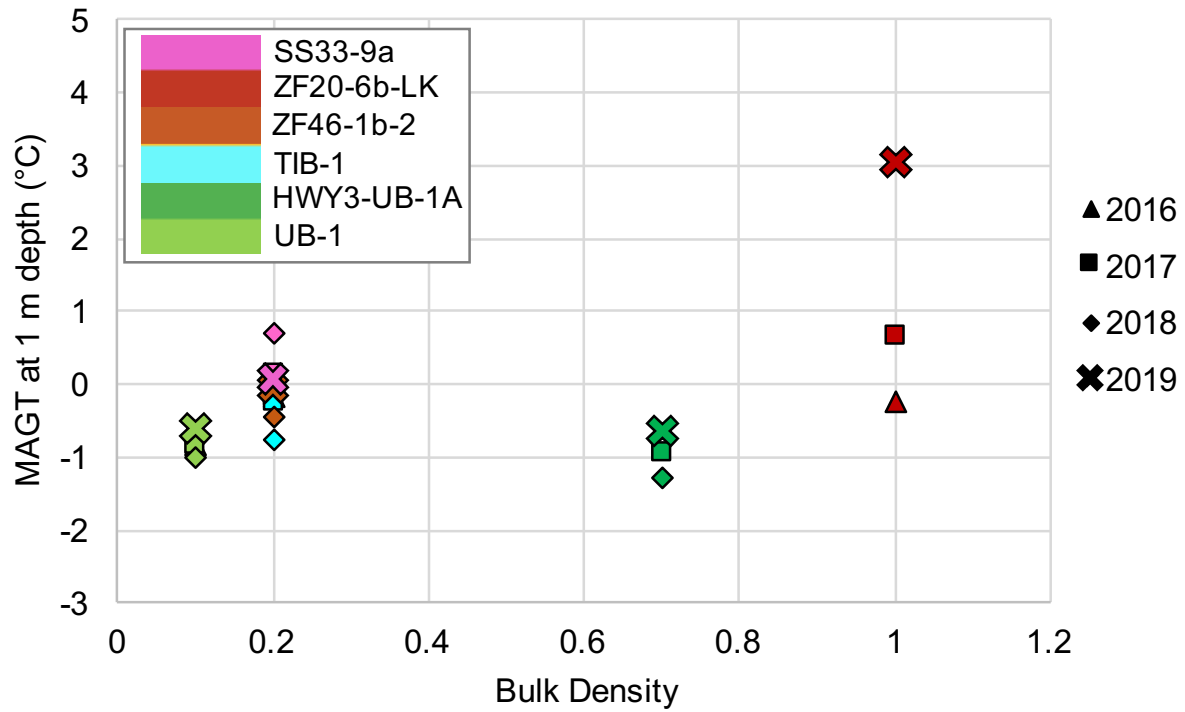


Figure C5: Changes in mean annual ground temperature (MAGT) at 1 m depth at the monitoring sites in relation to the average soil bulk density (n=6) at the permafrost monitoring sites. Note: a single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green.

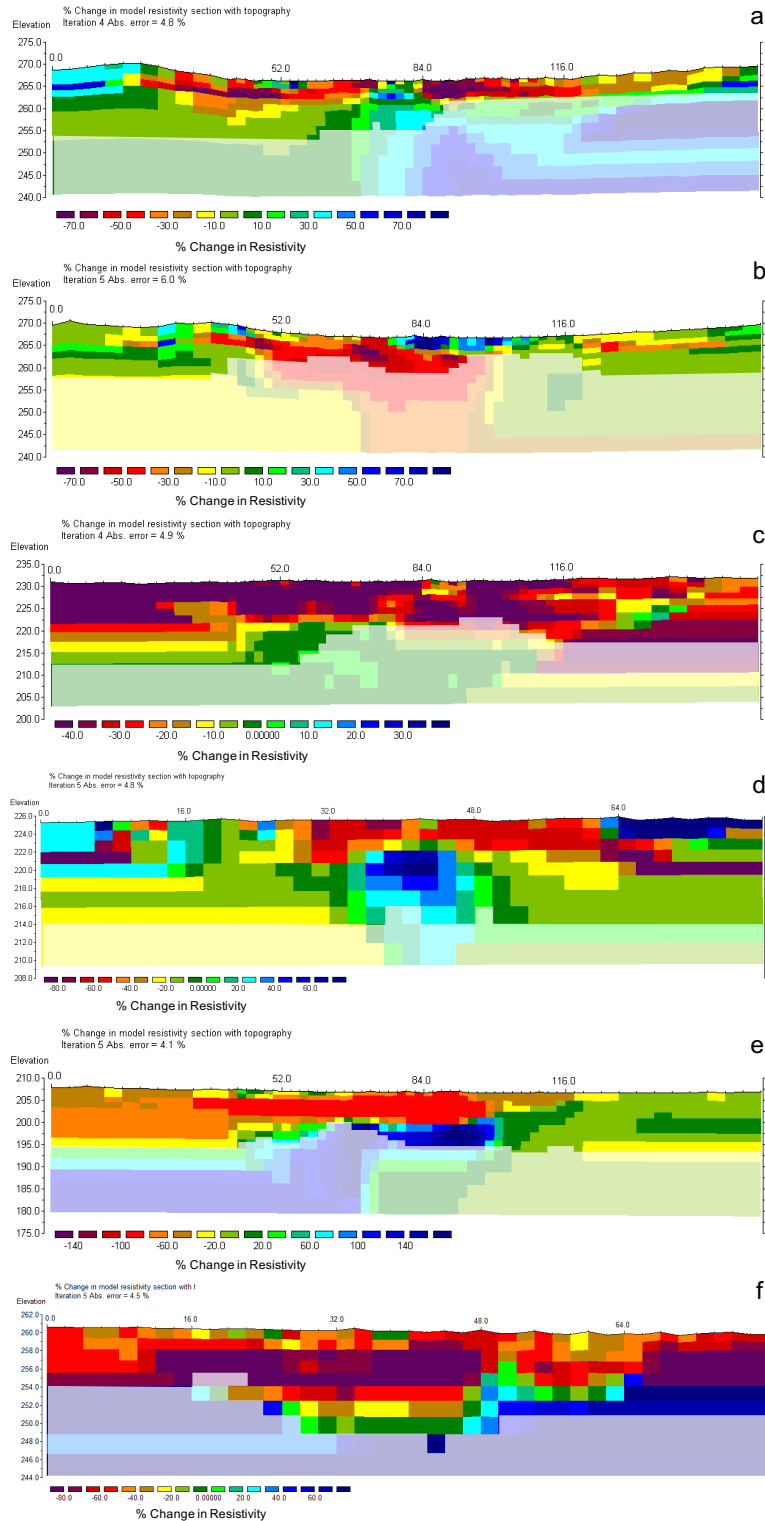


Figure C6: Time-lapse electrical resistivity profiles, presented as the percent (%) change in resistivity between 2015/16 and 2018 for (a) LL-PEAT, (b) LL-LITH2, (c) SS33-9a, (d) ZF46-1b-2, (e) ZF20-6b-LK, and (f) ZF46-1c. Site information is shown in Table C1. No time-lapse profile is available for site ZF46-18b.

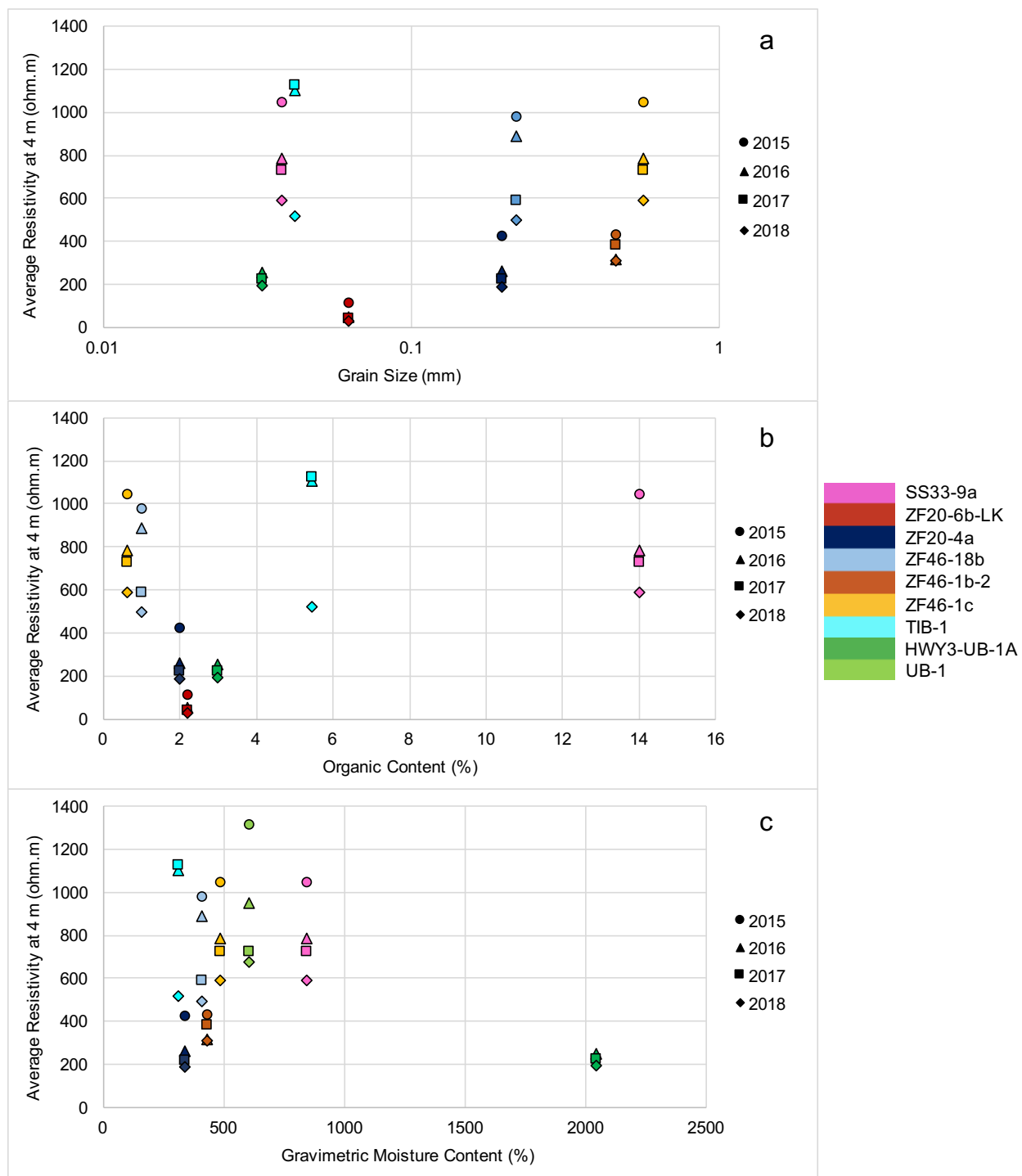


Figure C7: Average resistivity at 4 m depth in individual years compared to the (a) average grain size in the soil column, (b) average organic content, and (c) average gravimetric moisture content of organic soil in the active layer. Note: a single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green.

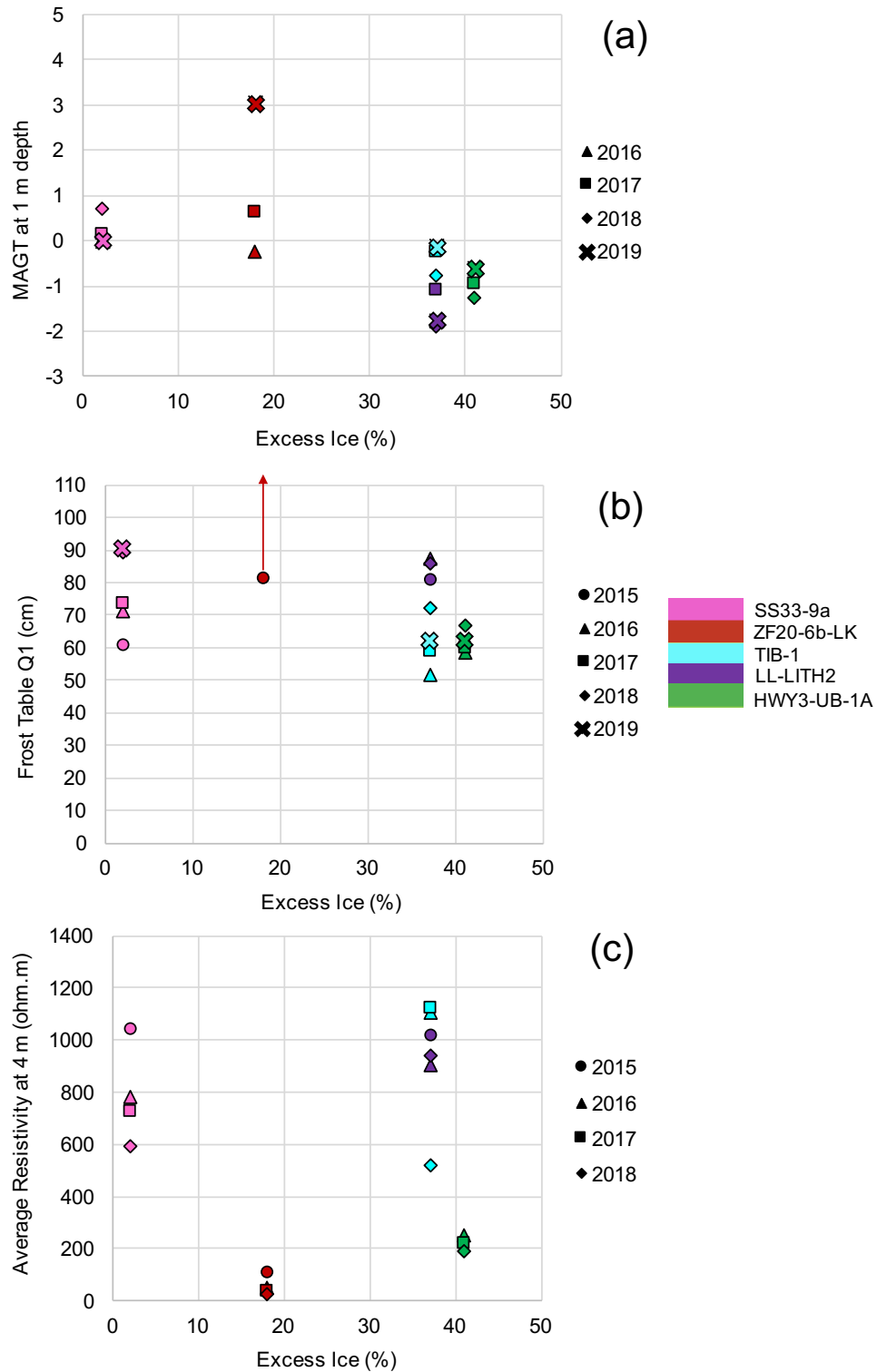


Figure C8: Associations between excess ice content (where present) and changes in (a) the surface offset, (b), the first quartile (Q1) of frost table depths, and (c), and the average resistivity at 4 m depth. Note: arrows indicate that measurements in later years were greater than the length of the probe; a single colour is used for each site, organized south (top) to north (bottom) on the legend, with unburned sites in green.

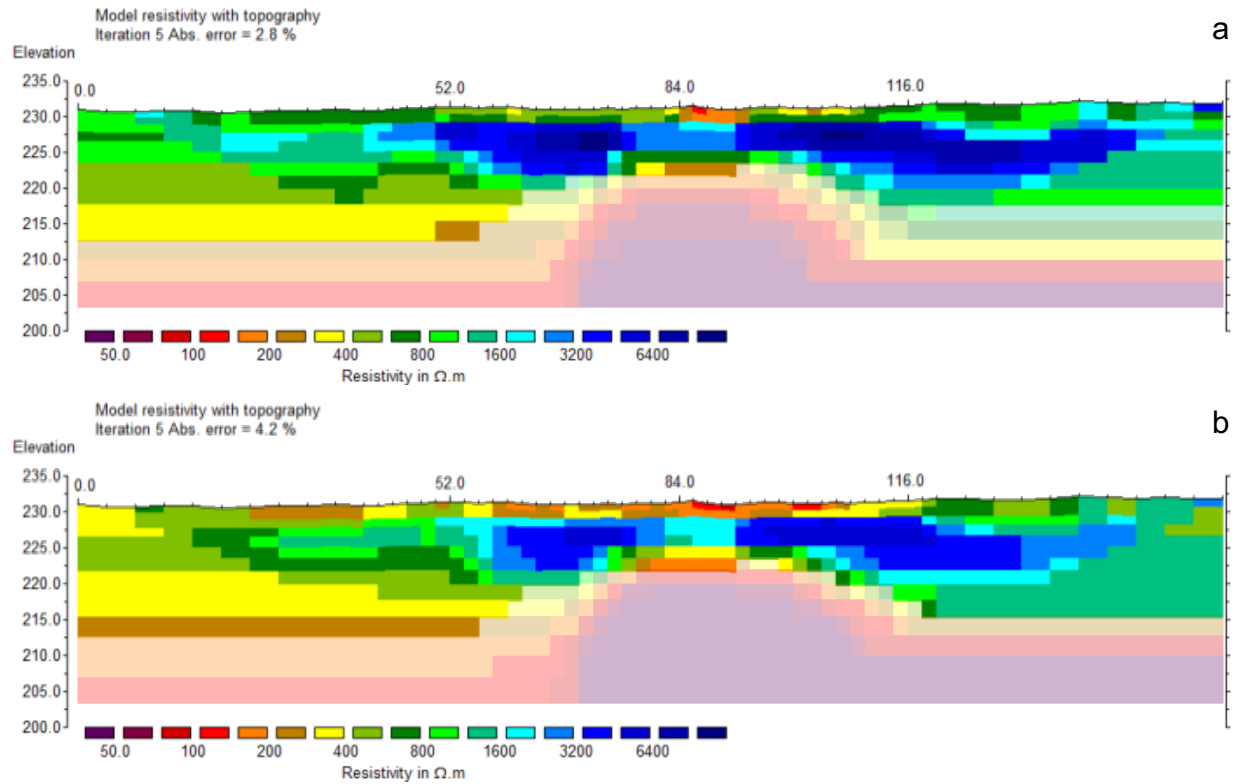


Figure C9: ERT surveys for site SS33-9a in (a) September 2015 and (b) September 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded.

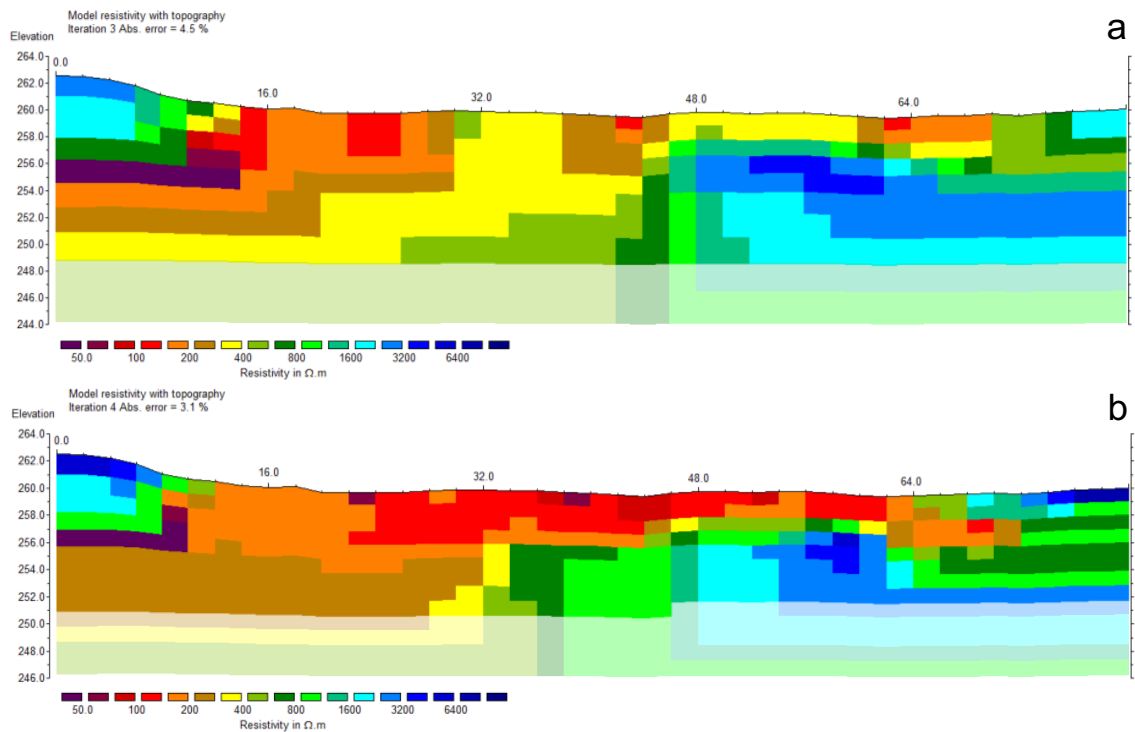


Figure C10: ERT surveys for site ZF46-1b-2 in (a) September 2015 and (b) September 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded.

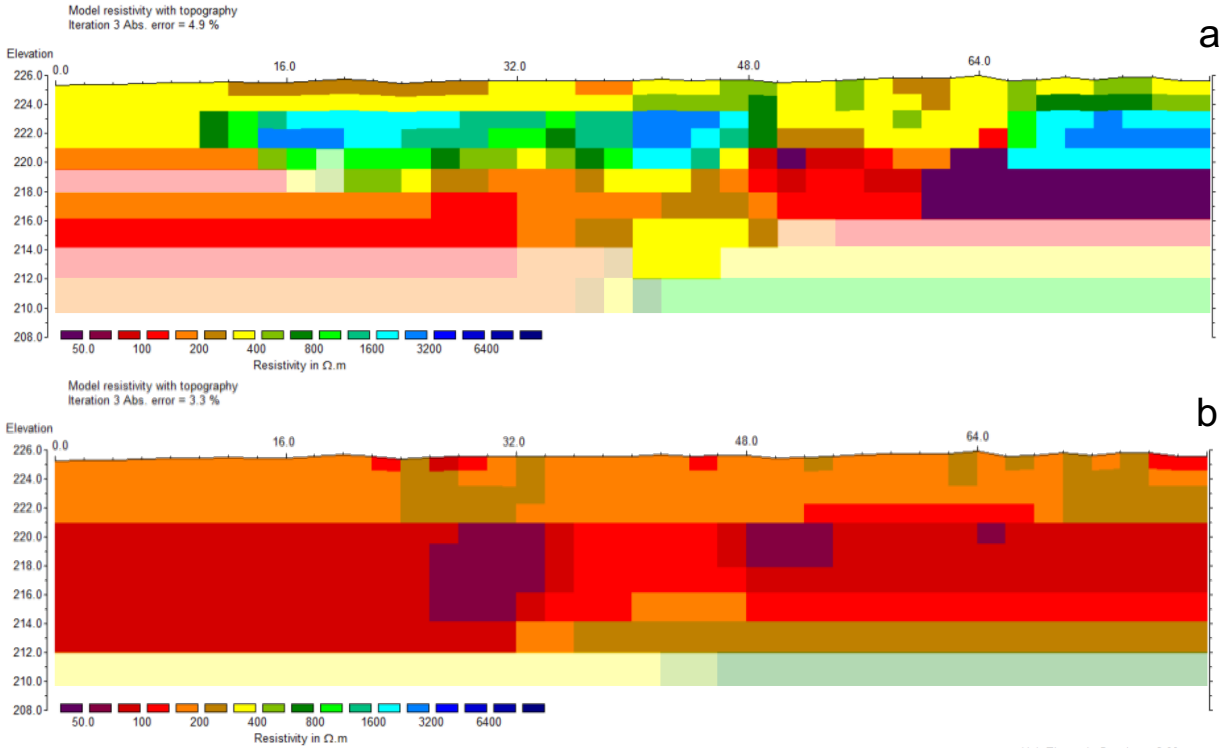


Figure C11: ERT surveys for site ZF46-1c in (a) September 2015 and (b) September 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded.

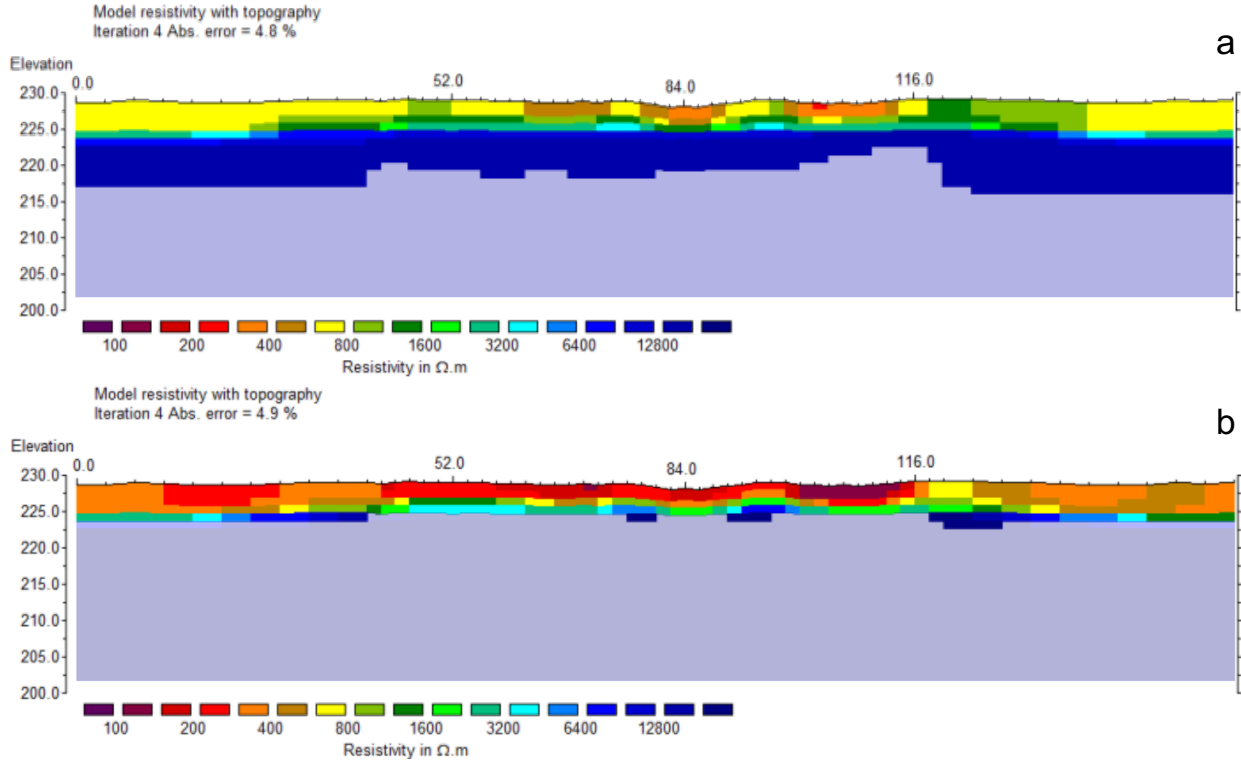


Figure C11: ERT surveys for site TIB-1 in (a) September 2016 and (b) September 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded.

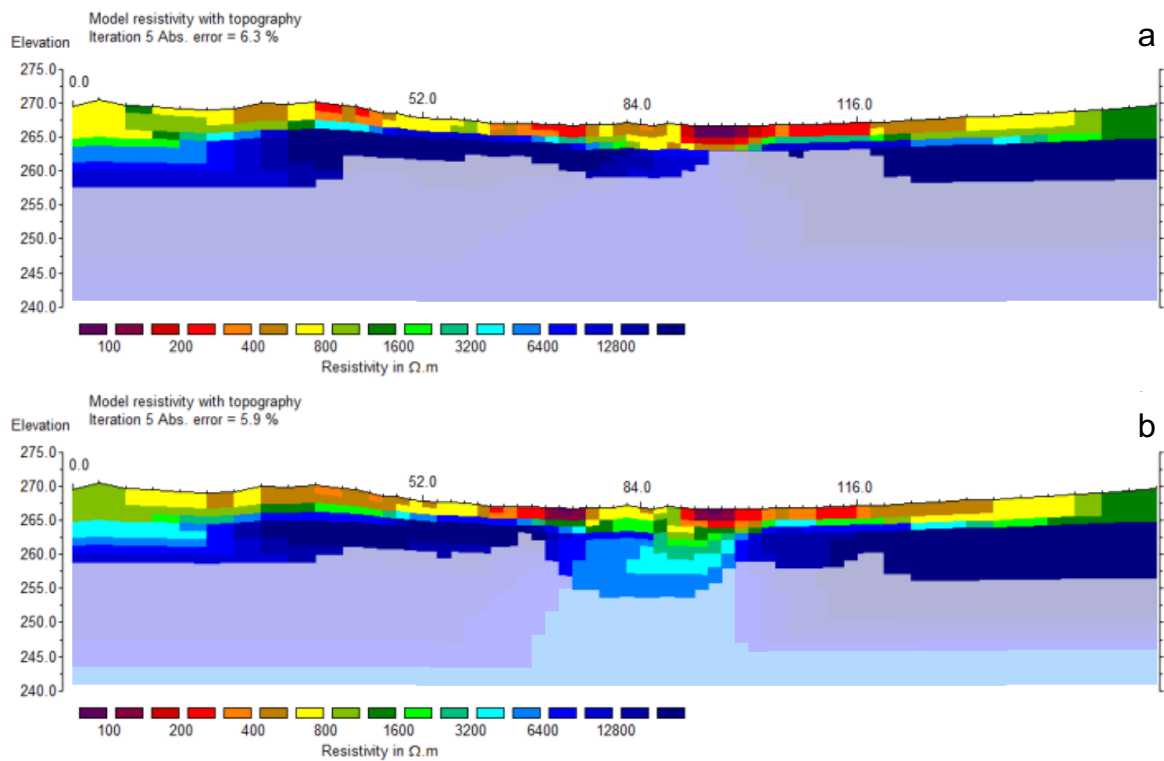


Figure C12: ERT surveys for site LL-LITH2 in (a) September 2015 and (b) September 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded.

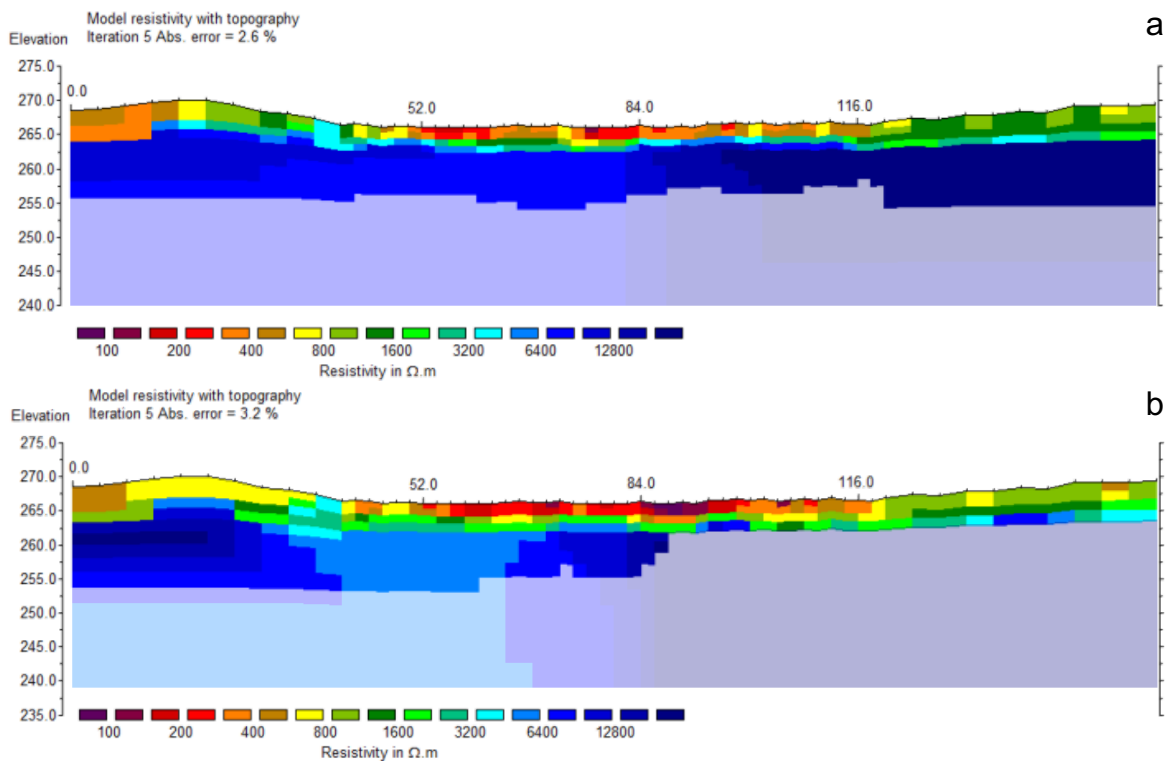


Figure C13: ERT surveys for site LL-PEAT in (a) September 2015 and (b) September 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded.

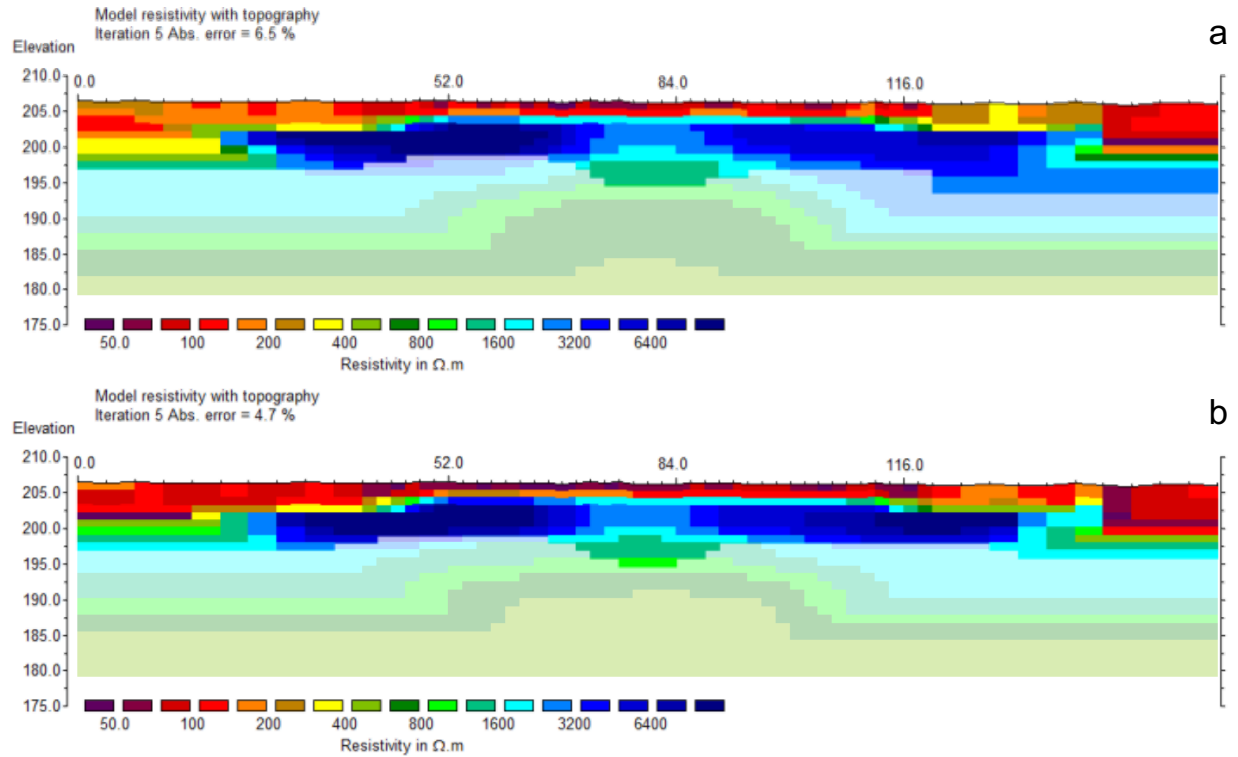


Figure C14: ERT surveys for site HWY3-UB-1A in (a) September 2016 and (b) September 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded.

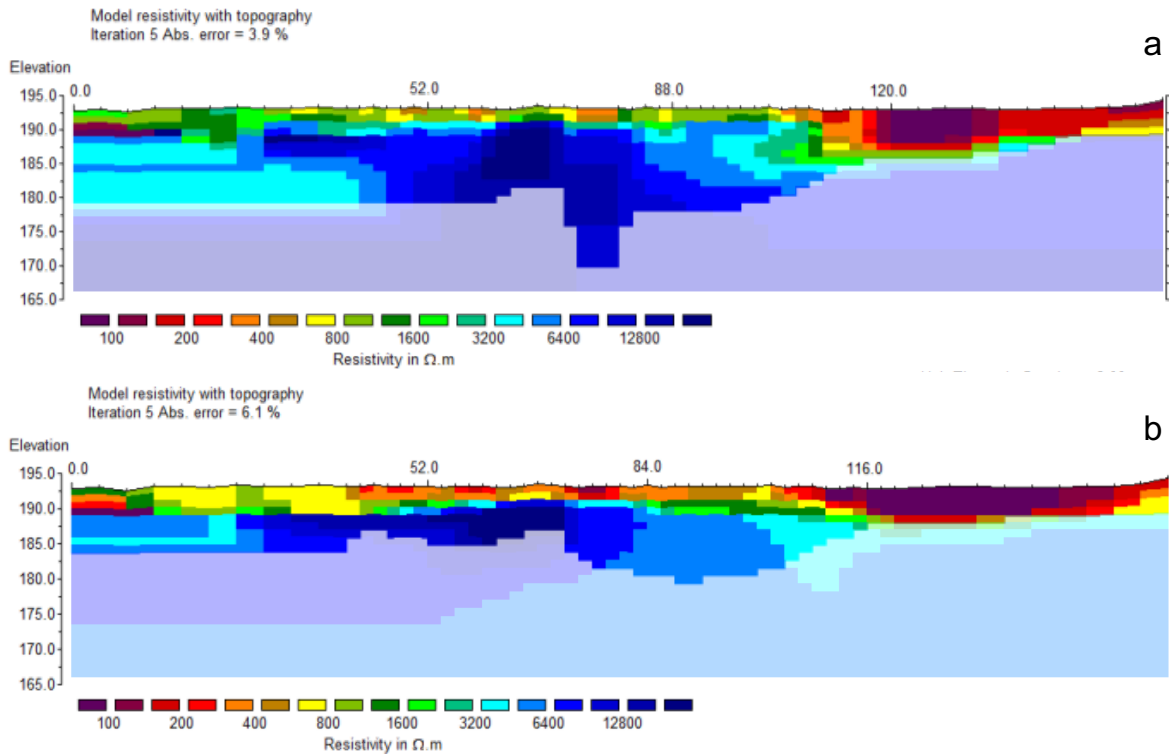


Figure C15: ERT surveys for site UB-1 in (a) September 2015 and (b) September 2018. ERT surveys are shown as model blocks, and those with sensitivity values less than 0.1 are faded.