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Modeling permafrost stability in peatlands with climate change and disturbance

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MODELING PERMAFROST STABILITY IN PEATLANDS WITH CLIMATE CHANGE
AND DISTURBANCE

BY

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Bachelor Degree of Arts, Mount Holyoke College, 2005

THESIS

Submitted to the University of New Hampshire
In Partial Fulfilment of
The Requirements for the Degree of

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In
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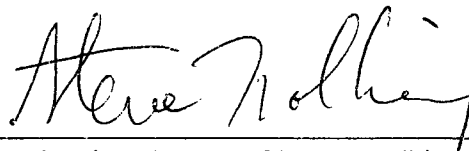
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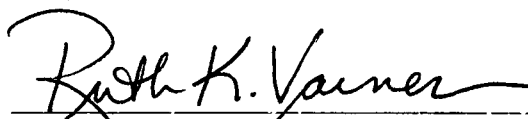


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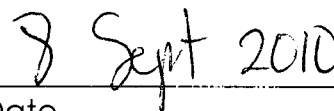
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ABSTRACT

MODELING PERMAFROST STABILITY IN PEATLANDS WITH CLIMATE CHANGE AND DISTURBANCE

by

Claire Treat

University of New Hampshire, September, 2010

Boreal and arctic regions are predicted to warm faster and more severely than temperate latitudes. They contain large stocks of below-ground soil carbon in peatlands and frozen soil, and the flux of the soil C to the atmosphere may be a strong feedback to climate change.

I compared the effects of climate change and wildfire on permafrost in peatlands using a soil thermal model. The model simulates soil temperatures and active layer thickness. I evaluated the model at a sedge-dominated Canadian arctic fen. I estimated the sensitivity of permafrost to current temperatures, future temperature projections, and wildfire.

Increases in air temperature due to climate change will increase surface soil temperatures, soil temperatures at depth, active layer depths, and growing season length, but not degrade permafrost by 2100 at this

site. Both wildfire and climate change increase active layer depths by 25 cm, but effects of wildfire diminish following vegetation recovery.

CHAPTER I

INTRODUCTION

Overview: Permafrost, soil carbon, and climate change

High latitudes are experiencing effects of climate change including degrading permafrost and altered hydrology due to warmer temperatures (Anisimov *et al.* 2007; Chapin *et al.* 2005; Euskirchen *et al.* 2006; Hinzman *et al.* 2005; Serreze *et al.* 2000). Warmer temperatures have lead to wide-spread increases in soil temperatures, and, subsequently, the degradation of permafrost has been observed across Alaska (Jorgenson *et al.* 2006; Osterkamp 2005; Osterkamp *et al.* 2009), Siberia (Smith *et al.* 2005), and boreal Canada (Beilman *et al.* 2001; Camill 2005; Halsey *et al.* 1995; Thie 1974; Vitt *et al.* 2000; Zhang *et al.* 2008).

The effects of permafrost degradation, thaw, and collapse on ecosystem processes have been an area of much recent research due to the implications for C storage in permafrost soils (Schuur *et al.* 2008). An estimated 50% of below-ground soil carbon (C) globally (~1650 Pg carbon) is stored in the northern circumpolar permafrost region, over 80% of which is in areas with permafrost (Tarnocai *et al.* 2009). Thick Organic soils (peatlands) store ~30% of SOC found in the permafrost zone (Tarnocai

et al. 2009), a proportionally large amount of C given their area (20% of the permafrost area).

Permafrost degradation has been predicted to continue into the future resulting in a 20 - 85% decrease in near-surface permafrost area by 2100 (Anisimov and Nelson 1996; Lawrence *et al.* 2008; Zhang *et al.* 2008). The projections of changes in permafrost area and the observed ecosystem responses have led numerous researchers to report the potential for massive C release due to increased temperatures and permafrost thaw from both carbon-rich mineral (Schuur *et al.* 2008; Schuur *et al.* 2009; Tarnocai *et al.* 2009; Walter *et al.* 2006; Zimov *et al.* 2006) and organic soils (Ise *et al.* 2008).

Implications Of Permafrost Degradation

Causes of permafrost degradation. Permafrost degradation can be caused by several different factors: changes in temperature, snowpack, or disturbance of the surface organic layer. Permafrost degradation has occurred in response to increases in air temperatures (Kokfelt *et al.* 2009; Osterkamp 2005; Romanovsky *et al.* 2007; Smith *et al.* 2005). Osterkamp (2005) found larger increases in permafrost soil temperatures during the winter than the summer, but increased air temperatures resulted in no change of the active layer depth (the maximum depth of summer thaw),

only warmer soil temperatures. Permafrost degradation has also been caused by changes in snowpack, either from increased snowfall (Christensen *et al.* 2004; Payette *et al.* 2004; Stieglitz *et al.* 2003) or from increased shrub density (Sturm *et al.* 2005). Shrubs and permafrost thaw features capture snow, creating deeper snowpack and warmer soil temperatures (Sturm *et al.* 2005). Additionally, permafrost degradation can be caused by the removal or disturbance of the surface soil layer, such as after wildfire (e.g. Dyrness 1982; Hayhoe and Tarnocai 1993; Viereck *et al.* 2008; Zoltai 1993).

Pathways of permafrost degradation. Permafrost degradation can result in small or drastic changes in permafrost and ecosystems (Jorgenson and Osterkamp 2005). Small changes in permafrost include talik formation; drastic changes include thermokarst. If the soil doesn't completely re-freeze the active layer, a zone of perennially unfrozen soil develops above the permafrost called a talik (Davis 2001). Talik formation can result in the eventual disappearance of permafrost, especially if the permafrost is shallow (Zhang *et al.* 2008).

Thermokarst occurs when permafrost with high ice content thaws. Subsidence or collapse of the permafrost surface (including the vegetation), also known as thermokarst, occurs as the ice melts (Davis 2001). Often the thermokarst area becomes a wetland (Jorgenson *et al.* 2001; Thie 1974; Zoltai 1993), although drainage may also occur resulting in

drier soils (Yoshikawa and Hinzman 2003). Thermokarst has occurred in 42% of the landscape with permafrost in one lowland area of interior Alaska (Jorgenson *et al.* 2001).

Ecosystem responses to permafrost degradation. Ecological and hydrological changes accompany permafrost degradation and thermokarst. Talik formation can lead to increased hydrological conductivity (Yoshikawa and Hinzman 2003) as well as increased biological activity and increasing shrub abundance in former graminoid-dominated arctic tundra (Schuur *et al.* 2007; Sturm *et al.* 2005).

More dramatic changes accompany thermokarst. In peatland areas, thermokarst degradation has resulted in internal lawns, collapse scar bogs and fens (Beilman *et al.* 2001; Vitt *et al.* 1994). A species composition shift occurs as bog species (*Sphagnum* mosses, black spruce) are replaced by highly productive fen species (sedges and mosses species) (Camill 1999; Harris and Schmidt 1994; Robinson and Moore 2000; Turetsky *et al.* 2000). In forested areas, black-spruce or aspen-dominated forests have become fens with floating mat vegetation following permafrost collapse (Jorgenson *et al.* 2001; Osterkamp *et al.* 2000).

Peatland succession within permafrost areas is rapid (Payette *et al.* 2004). As peat accumulates in collapse areas, the peat surface becomes elevated relative to the water table and favors species better adapted to dry environments. Shrub invasion changes snow distribution across the

landscape through the creation of windbreaks subsequent snowpack reduction in areas without shrubs can lead to permafrost aggradation (Lawrence and Slater 2005; Robinson and Moore 2000; Zoltai 1993).

Similarly, permafrost aggradation has been observed in hummocks with low snowpack in Finnish Lapland (Seppala 1998). Rapid peat formation and subsequent insulation of existing permafrost was observed following permafrost degradation in the arctic (Jorgenson *et al.* 2006). Over longer time scales, work by Zoltai (1993) and Robinson and Moore (2000) demonstrated a cycle of permafrost degradation and collapse, followed by vegetation succession from fen species to bog species, and, after 100-200 years, the eventual aggradation of permafrost following the establishment of Sphagnum mosses.

Carbon response to permafrost degradation. Dramatic changes in ecosystems following permafrost degradation result in altered C cycling due to changes in soil temperatures, vegetation and productivity. Warmer soils result in increased rates of decomposition and C mineralization (Dorrepaal *et al.* 2009; Dutta *et al.* 2006; Turetsky *et al.* 2002). In previously frozen soils, radiocarbon dating in several studies has shown that C emissions are derived primarily from older soil C, indicating that soil C found in newly thawed soil is relatively labile (Dorrepaal *et al.* 2009; Dutta *et al.* 2006; Schuur *et al.* 2009).

Thermokarst results in warmer and generally wetter conditions. This has led to enhanced C sequestration in thermokarst areas as species composition changed and resulted in increased rates of C accumulation in peatlands (Camill *et al.* 2001; Harris and Schmidt 1994; Robinson and Moore 2000; Turetsky *et al.* 2000; Vitt *et al.* 2000) and in former forests (Myers-Smith *et al.* 2008; Myers-Smith *et al.* 2007). However, enhanced C uptake rates or C losses may not be sustained over longer time scales (Oechel *et al.* 2000; Robinson and Moore 2000). Sustained, elevated CO₂ emissions were observed at a sub-Arctic blanket bog for the duration of an 8-year temperature manipulation experiment (Dorrepaal *et al.* 2009).

Wetter and warmer soil conditions in thermokarst also result in increased methane emissions for both peatland (Bubier *et al.* 1995; Christensen *et al.* 2004; Turetsky *et al.* 2002), former black spruce forests (Wickland *et al.* 2006), and tundra ecosystems (Oechel *et al.* 1998). Interactions between experimental soil moisture and temperature manipulations resulted in substantially larger increases in CH₄ release from warmer and wetter soil than either warmer or wetter soils (Turetsky *et al.* 2008).

The net results of changes in C uptake and emissions is unclear. In some cases, no significant change is observed. Payette *et al.* (2004) observed very little net change in C status, as increased C emissions were offset by increased C storage by paludification of peatlands. In other

cases, the uncertainty of the C measurements is too large to determine the net carbon balance of the site (Wickland *et al.* 2006). In still other situations, similar expansion of peatland area and peatland succession resulted in increased C storage in interior Alaska (Myers-Smith *et al.* 2007).

Interactions Between Organic Soil And Permafrost

Local, site-specific conditions and thermal inertia have mediated the impacts of climate change, leading to the existence of permafrost at disequilibrium with the climate, even in areas where the mean annual temperature can be above freezing (e.g. Camill and Clark 1998; Halsey *et al.* 1995; Seppala 1998; Shur and Jorgenson 2007; Vitt *et al.* 1994). In particular, Sphagnum peat and Sphagnum moss species and insulate permafrost due to their especially low thermal conductivity when dry, resulting in much cooler soil temperatures (Camill and Clark 1998; Robinson and Moore 2000; Vitt *et al.* 1994). Sporadic and isolated patches of permafrost have persisted for several hundred years in disequilibrium with the climate within peatlands, due to the insulating effects of thick organic soils and mosses (Shur and Jorgenson 2007; Vitt *et al.* 2000). Permafrost has even reformed at some sites with favorable species composition, climatic conditions, and peat deposition rates (Camill 1999; Halsey *et al.* 1995; Vitt *et al.* 2000).

At a larger scale, Alexeev *et al.* (2007) and Yi *et al.* (2007) have demonstrated the importance of including an organic soil layer to accurately model permafrost. The inclusion of organic soil in global soil temperature models resulted in a 25% greater permafrost area than without inclusion of organic soil (Lawrence *et al.* 2008). Additionally, the depth of organic soils may be underestimated in areas because some permafrost models may distribute the organic soil from a small, deep peat deposit uniformly over the grid cell resulting in a larger area with shallower organic soil (e.g. Wania *et al.* 2009).

Disturbance Effects On Permafrost

Disturbance is a key factor that affects local soil thermal, biogeochemical, and ecosystem processes. Wildfire is a widespread disturbance within the boreal zone and is currently less common in the arctic (Stocks *et al.* 2002). Wildfire is estimated to release an average of 105.9 – 208.5 Tg C yr⁻¹ in the boreal zone, including Canada, Alaska (US) and Russia (Kasischke *et al.* 2005); in recent decades, the frequency of wildfire has increased, in addition to larger areas being burned (Kasischke and Turetsky 2006). A continued increase in frequency and burn area is likely due to increases in surface temperatures, growing season length, and changes in soil moisture (Flannigan *et al.* 2005).

Disturbance that removes part of the surface organic layer alters the soil thermal regime (Hayhoe and Tarnocai 1993; Zoltai 1993). Examples of disturbance include both anthropogenic and natural activities such as logging, development, peat harvesting, wildfire, and fire suppression methods. A long-term study of the effects of surface organic soil removal showed a persistent increase in depth to permafrost after 36 years, with differences observed between sites with mechanical removal and wildfire (Dyrness 1982; Viereck *et al.* 2008).

Permafrost degradation in peatlands is often triggered by disturbance (Zoltai 1993), although disturbance does not necessarily lead to permafrost degradation (Sannel and Kuhry 2008), especially if the organic horizon is relatively thick. For example, the soil temperature at 15 cm depth did not increase during wildfire in a site with thick organic soil (Yoshikawa *et al.* 2002).

At a local scale, permafrost stability is affected not only by species composition and other local conditions, but also by the frequency of disturbance, severity of disturbance, and the subsequent removal of organic soil (Camill and Clark 1998). At a regional scale, permafrost stability will be affected by changes in climate (temperature, precipitation, and seasonality), hydrology, and vegetation dynamics (Camill and Clark 1998).

The relative magnitude of climate change and disturbance impacts on permafrost stability in peatlands have not been directly compared. In this paper, I adapt an existing permafrost model to peatlands by including an organic soil column and dynamic soil moisture. I examine the sensitivity of the permafrost model to changes in soil moisture, thermal properties and organic layer thickness. Finally, I compare permafrost stability given future climate scenarios and changes in disturbance regimes. The relative impacts of these two types of disturbance have significant implications for the growing season length and seasonality, hydrology, and subsequently the carbon balance of peatland sites experiencing permafrost degradation and altered disturbance regimes.

CHAPTER II

METHODS

Model Description

I used the Geophysical Institute Permafrost Lab (GIPL) 2.0 permafrost model previously described by *Marchenko et al.* (2008). The GIPL-2.0 model solves non-linear heat transfer with phase change in one dimension numerically. I simulated soil temperature dynamics from 0 to 100 meters on a daily time step; analysis of results led to a simulated depth of seasonal thaw. Daily surface soil temperatures, water table level, and a description of the underlying soil and bedrock were used as input for the GIPL model. Soil moisture, thermal conductivity, and heat capacity values from *Marchenko et al.* [2008] were used for bedrock (depths greater than 10 m).

To adapt the model to peatland ecosystems, I added a layered soil column and dynamic soil moisture. The layered soil column had 5 soil layers: three peat layers, a mineral soil layer, and a bedrock layer (Fig. 1). The soil properties of the peat layers (horizon thickness, porosity, and water retention) and organic content and composition of the mineral soil layer are both depth- and site- specific (Table 1). Three peat layers were used to accurately capture the changes in bulk density and water retention

characteristics at depth in peat. The thermal conductivity and heat capacity of peat and mineral soils were calculated using methods described by *Wania et al. (2009)* and *Granberg et al. (1999)*. I used values of thermal conductivity and heat capacity from *van Wijk and de Vries [1963]* and a regression to determine the heat capacity of peat from the volumetric water content (Table 1). I included calibration coefficients, C_f and C_t , as multipliers of the frozen and thawed soil thermal conductivity, respectively.

Accurately capturing the soil moisture distribution in peat soils is important for modeling soil thermal dynamics due to the insulating effects of dry sphagnum at the peat surface (*Vitt et al. 2000*; *Waelbroeck 1993*). Rather than use a soil profile with constant soil moisture at all depths, I used observations of water table to determine the soil moisture (Fig. 1). I assumed saturation in soils below the water table. The water content in soils above the water table was a function of height above the water table and peat type (Table 1) due to differences in water holding capacity that arise from differences in peat porosity and bulk density (*Weiss et al. 2006*).

To calculate the water content in peat above the water table, the water table measurements were converted to depth-distributed percent of water filled pore space (WFPS) using empirical relationships for peat soil developed at Mer Bleue Bog (*Frolking et al. 2002*). The slope of the

relationship between water table position and WFPS differed by peat horizon (Table 1). The WFPS was multiplied by porosity to determine the volumetric water content, which was used in calculations for thermal conductivity and heat capacity.

Site Description

I used observations of surface soil temperature, water table level, soil moisture, and soil properties from Daring Lake Fen in the Daring Lake Research Area, Northwest Territories, Canada (64.865° N, 111.567° W) for model validation. Daring Lake Fen is an Arctic Fen with permafrost, dominated by *Carex* spp. with ~60 cm of peat underlain by silt loam (Lafleur and Humphreys 2007). Mean annual air temperature from 2006 to 2008 was -8.7°C; the mean monthly minimum and maximum temperatures were -32°C and -27°C in January and 8°C and 18°C in July, respectively. For 2006 to 2008, the mean total precipitation from 15 May - 31 August was 104 mm.

At Daring Lake Fen for 2006 to 2008 there were daily measurements of air temperature, soil temperature (0, 2, 5, 10, 20, 40 and 60 cm depths), snow depth, and water table level (2007, 2008 growing season only). Surface soil temperature data were used as model input; gaps were filled using a linear regression with soil temperature at 2 cm depth. Water table depth during the remainder of the year (i.e. non-growing season) was

held constant from freeze-up to snowmelt, as determined by snow depth data. Following snowmelt, the water table increased incrementally until the first water table measurements were recorded. There were no data for the water table position during the 2006 growing season; 2006 water table position was estimated by using the mean water table position of the 2007-08 growing seasons and was held constant throughout the growing season.

Statistical Analysis

To evaluate the model predictions of soil temperatures at Daring Lake Fen, I compared the predictions to observed soil temperature data. To quantify the agreement between modeled and observed soil temperatures, I used the Root Mean Squared Error (RMSE) at different depths and the differences between the number of modeled and observed thaw days at 40 cm and 60 cm.

To find the RMSE, I used a linear model of the predicted soil temperatures against the observed temperatures at 20 cm, 40 cm, and 60 cm depths. To calculate the number of thaw days at depth, I summed the number of days per year with mean temperatures greater than 0 °C. My final model selection was based on a combination of lowest RMSE of modeled temperatures across depths and smallest differences between the number of modeled and observed thaw days at 40 cm and 60 cm depths (Table 2).

To evaluate the differences between treatments in the sensitivity and climate change analyses, I used the analysis of variance procedure with post-hoc analysis. I used the Tukey 'Honest Significant Difference' post-hoc analysis to determine confidence intervals for sample means and determine the corrected differences between means for multiple sample groups. All statistical analyses were done using R Statistical Software (R Development Core Team, 2008). Additionally, I used the Zoo package for R (Zeileis and Grothendieck 2005) to calculate rolling means of environmental variables and active layer depths over 5-year time scales to depict longer-term trends.

CHAPTER III

MODEL EXPERIMENTS AND SENSITIVITY ANALYSES

I predicted the future stability of permafrost with climate change, and compared the relative effects of climate change and disturbance by fire. Additionally, I evaluated the effects of organic soil thickness, the timing of wet and dry periods, and the distribution of soil moisture within the peat column: namely, whether a water table was present or there was constant soil moisture with depth. My hypotheses were:

1. Thicker organic soils will experience less permafrost degradation due to lower thermal conductivity and higher soil moisture contents. Higher soil moisture also increases the heat capacity, resulting in a larger heat sink.
2. Including a water table will decrease active layer depths due to lower total thermal conductivity at the dry peat surface.
3. The timing of wet and dry periods will affect the transfer or loss of heat to deeper peat depths through higher thermal conductivity.

4. More severe burns will increase soil temperatures and active layer depths due to the removal of insulating peat by wildfire.

5. Burn timing will affect active layer depths through increased burn severity. Kasischke and Johnstone (2005) found a difference in burn severity between burns that occurred early in the growing season and later due to drier soils. They attributed drying to seasonal soil thawing as the depth to permafrost increased and allowed for increased soil drainage.

Climate Change

Future scenarios simulations are from the ECHAM 5 model for the IPCC assessment report #4 (Roeckner *et al.* 2003). I used two data sets from ECHAM5: the 20C3M, which are the 20th century control runs that use measured CO₂ from 1900-2000 and assume constant CO₂ emission for the future (2001 – 2100), and SREA1B, the data for the IPCC scenario A1B (2001-2100), which assume rapid economic growth, introduction of new, more efficient technology, a peak in population, and reliance on multiple energy sources. I extracted daily values of air temperature, surface soil temperature (3 cm), precipitation, and snow water equivalent for the grid cell containing Daring Lake Fen. As considerable uncertainty arises from

scaling $0.5^{\circ} \times 0.5^{\circ}$ grid-cell climate predictions to a point, I scaled ECHAM5 precipitation predictions to observed climate normals to reduce bias (Table 3).

ECHAM5 modeled air and soil temperatures compared well with observed air and soil temperatures at Daring Lake Fen (Table 3) for the summer months, while there was a bias towards colder air and soil temperatures in ECHAM5 during the winter months.

A direct comparison of modeled precipitation from ECHAM5 with observations at Daring Lake Fen was more difficult. Daily precipitation and snow water equivalents were extracted from ECHAM5 20c3 scenarios. Daily precipitation observations at Daring Lake Fen were available from 15 May – 31 August of 1997-2005. Snow depths at Daring Lake Fen were measured for 2006-2008 using radiation balance methods [E. Humphreys, pers. comm.]. To better compare the ECHAM5 20c3 predictions to observations from Daring Lake Fen, I compared observed data at Daring Lake Fen and predictions from ECHAM5 20c3 climate scenarios to observations at Lupin A climate station, located 100 km north of Daring Lake Fen ($65^{\circ} 45.600'$ N, $111^{\circ} 15.000'$ W; <http://climate.weatheroffice.gc.ca/>).

A comparison between Daring Lake Fen and Lupin A climate station showed slightly lower mean summer precipitation at Daring Lake Fen than at Lupin A (104 mm at Daring Lake Fen vs. 131 mm at Lupin A;

Table 3). ECHAM5 modeled precipitation for June through August was nearly double the mean observed precipitation at Daring Lake Fen from 1997-2005 (Table 3; Lafleur and Humphreys 2007). Similarly, differences in annual precipitation between Lupin A and ECHAM5 scenarios were large; mean annual precipitation at Lupin A from 1971-2000 was 299.2 mm, compared with 487.3 mm for 20c3 during the same period and 490.8 mm for A1b from 2001-2030 (Table 3).

I calculated water table and soil moisture from the precipitation data using scaling relationships by correlating the distribution of climate normal ECHAM5 precipitation values with the observed water table distribution. From 15 May 2007 – 31 Aug. 2007, Daring Lake Fen received 60.2 mm of precipitation, which was on the dry end. The lowest measured precipitation at Daring Lake Fen for this period from 1997-2005 was 52.3 mm, while for the same period in 2008, precipitation at Daring Lake Fen was 184.4 mm (maximum measured seasonal precipitation for all years was 196.0 mm). In scaling, I assumed the range of water table observed adequately captured the water table distribution and that total summer precipitation was correlated with total annual precipitation at Daring Lake Fen, as it was at nearby Lupin A climate station ($b = 1.09$, $B_0 = 131.6$; $p = 1.453 \times 10^{-7}$, $F(1,23) = 55.36$, $r^2 = 0.7065$). The scaling relationship that I developed between ECHAM5 annual precipitation values and mean annual water table at Daring Lake Fen was:

$$WT = -46.98 + 7.937 \times 10^{-2} * P \quad [1]$$

where WT is the mean annual water table position in centimeters (negative values indicated a water table below the surface) and P is the annual precipitation given by the ECHAM5 model in mm/yr. The mean annual water table position was used as the daily water table position for each year. We tested sensitivity to the timing of wet and dry periods in separate simulations (Section 3.3). Water table pooling at the surface was considered to be unlikely due to good site drainage [E. Humphreys, pers. comm.], so scaled water tables were truncated at the surface.

I determined the response of soil temperature and active layer depth at Daring Lake Fen to three future climate scenarios (Table 4; Fig. 3). The first scenario, 20c3, was used to recreate soil temperatures from 1900-2001 and predict soil temperatures from 2001-2100 with no change in air temperature or precipitation relative to the 1900-2000 (Fig. 3). The second future scenario, A1b+T, used higher 21st century temperatures (ECHAM5 A1b surface soil temperature) and no change in precipitation (ECHAM5 20c3 precipitation). The third future scenario, A1b+T+WT, used higher 21st century temperatures (ECHAM5 A1b surface soil temperature) and a shallower water table in response to increased precipitation (ECHAM5 A1b precipitation). Both A1b scenarios ran for 2001-2100 and started from 20c3 soil temperature values in 2000 (Table 4).

Sensitivity To Water Table

The differences in soil moisture distribution affect thermal properties; dry peat at the surface may provide sufficient insulation to protect permafrost in areas where mean annual temperatures are greater than 0 °C (Vitt *et al.* 2000). I conducted a sensitivity analysis to determine the importance of including a dynamic water table in the model, as opposed to constant soil moisture throughout the soil profile and constant soil moisture throughout the growing season. With the inclusion of a water table, peat below the water table was saturated and the soil moisture content above the water table was determined by water retention curves (Table 1). With constant soil moisture, the soil moisture remained constant with depth and over time. The three treatments were 20%, 60%, and 100% saturation.

I ran the water table sensitivity scenarios using the A1b +T +WT climate for 100 years to determine the differences between inclusion of a water table and using constant soil moisture for a permafrost peatland. In the sensitivity analysis, I included the comparison between 20%, 60%, 100% constant soil moisture and using a daily water table equal to the mean annual water table.

Additionally, I evaluated the sensitivity of permafrost to climate change given different soil moisture conditions seasonally and over time. I

created a 1 mm yr⁻¹ increase and decrease in mean annual water table position relative to A1b +T +WT representing wetter and drier soils. I also evaluated the response of active layers to changes in moisture seasonality by using four different scenarios. For these scenarios, I defined spring as the months of April, May and June, and the fall months as October, November, and December. The growing season was from April through September, while the non-growing season was from October through March. I used a wet spring/dry fall scenario, where the water table was 10 cm higher and lower than the mean annual water table level for each year for the wetter and drier seasons, respectively. The mean water table changed by <3 mm in this scenario. I repeated this analysis for a dry spring/wet fall scenario, a dry growing season/wet winter scenario, and a wet growing season/dry winter scenario.

Sensitivity To Peat/Organic Soil Thickness

I evaluated the effects of organic soil (peat) layer depths on permafrost active layer depths and soil temperatures by altering peat thickness and the distribution of soil moisture. I reduced peat thickness from 60 cm (observed at Daring Lake Fen) to shallower depths ranging from 1 cm to 30 cm and increased the mineral soil thickness. For most shallower peat depths, I assumed the soil moisture of the mineral soil was

at field capacity (60% saturation) due to the correlation between poor to very poorly drained soil classes (i.e. water table) and high soil carbon storage (i.e. thick organic soils) found in boreal Canada (Harden *et al.* 2001). In these scenarios, I also assumed the peat was 60% saturated. For peat depths greater than or equal to 15 cm, the water table for the scenario was used and the water content in the unsaturated zone was a function of height above the water table (Section 2.1).

I used two analyses to determine the effects of organic layer thickness on the active layer depth of permafrost to elicit both long-term and shorter-term effects of organic soil thickness. The first approach compared the effects of different thicknesses of organic soil (5 cm, 15 cm, 30 cm, and 60 cm of peat) on annual active layer depths for the climate scenarios (control/20c3 and A1b +T +WT) from 1900 – 2100. This approach used the water table from each scenario to determine soil moisture for all peat depths rather than field capacity.

The second approach evaluated the effects of climate change on permafrost when different organic soil thicknesses were used. Organic soil thicknesses ranged from 1 cm to 60 cm. I compared active layer thicknesses among different organic soil thicknesses and soil moisture distributions (field capacity vs. water table) using climate change scenario A1b +T +WT from 2000-2100. This indicated the resulting

differences in active layer depths that were due to organic soil thickness rather than climate change.

Disturbance

I conducted two disturbance analyses with different purposes: (1) a sensitivity analysis to determine which factors significantly affect permafrost active layer depths post-fire, and (2) a long-term analysis to determine the impact of disturbance and climate change on permafrost and soil temperatures from 2001 – 2100. While I explicitly considered wildfire, my results are also applicable to other types of disturbance that remove organic soil and alter soil thermal properties.

In both analyses, I simulated the effects of wildfire in the same way. I removed the surface peat to the burn depth immediately following the day of the wildfire. This resulted in a new peat surface layer with the soil properties (bulk density, porosity, water holding capacity, and thermal properties) and soil moisture of a deeper peat layer immediately below the burn depth. After the simulated burn, the water table position was closer to the new post-burn surface by the amount of peat lost in the burn (the burn depth). Additionally, surface soil temperatures were altered following the wildfire to account for decreased shading and changes in albedo that accompany fire (Chambers *et al.* 2005; Chambers and Chapin 2002; Harden *et al.* 2006; Liu *et al.* 2005; O'Neill *et al.* 2002;

Randerson *et al.* 2006). I assumed no surface temperature increases after vegetation recovery, an estimated 21 years post-fire (Kuhry 1994; Wieder *et al.* 2009; Zoltai *et al.* 1998). Changes to the surface soil temperature were calculated on a daily time step and added to the surface soil temperature inputs to the model. Also, wildfire alters soil moisture due to both the removal of organic soils and decreases in evapotranspiration following vegetation die-off in a fire, resulting in wetter conditions in the surface soil (Johnstone and Chapin 2006; O'Donnell *et al.* 2009) and potentially drier conditions in the mineral soil (Harden *et al.* 2006). In all simulations, I held water table constant relative to the original peat surface, which meant that the water table was closer to (or above) the new, post-fire peat surface for the remainder of the simulation.

Sensitivity analysis: disturbance characteristics that significantly affect active layer depth. A sensitivity analysis allowed us to examine the importance of factors or treatments that I hypothesized would affect active layer depths and permafrost stability: burn depth (a function of fire severity), timing of wildfire within the season, and post-fire surface temperature response. The sensitivity analysis was run for the 10 years post-fire using soil temperatures and water table levels from climate scenario A1b +T +WT. Active layer depths were compared between treatments using analysis of variance.

To examine the impacts of burn severity on peatland soil temperatures and permafrost stability, I removed varying proportions of the organic layer thickness depending on the fire severity (Bonan 1990). I removed 5 cm of peat to simulate a shallow burn and removed 20 cm of peat to simulate a severe burn (Kasischke and Johnstone 2005). I also tested whether the seasonality of the fire affected the active layer depth by comparing burn timing; the early season burn occurred on 27 June, while the late season burn occurred on 23 September.

Multiple studies in the boreal region have found an increase in surface soil temperature after wildfire, ranging from 0.3°C to 13° C across various timescales and measurement techniques (Harden *et al.* 2006; Liu *et al.* 2005; Nakano *et al.* 2006; Wieder *et al.* 2009; Yoshikawa *et al.* 2002). Most of these studies were located in organic soils with a tree canopy. However, other researchers have found a 2°C – 3°C decrease in surface soil temperatures in burned peatland plots compared with paired unburned peatland plots in the years following a wildfire (Chambers *et al.* 2005; O'Donnell *et al.* 2009). This may be explained by increase in albedo due to the species composition shift from mosses to forbs and graminoids post-fire. An additional factor determining the post-fire soil temperature response may be the presence of trees; Daring Lake Fen is not a treed site, so the soil temperature response may more closely follow the results for a peatland (O'Donnell *et al.* 2009) or tundra (Chambers *et al.* 2005).

However, I did consider both potential responses of surface soil temperatures.

To capture post-fire burn temperature changes in the disturbance sensitivity analysis, I used the following equation:

$$\Delta T_w = 5.413e^{-5.62 \times 10^{-4} t_{burn}} - 0.25 \quad (\text{warming}) \quad [2]$$

$$\Delta T_c = -3.267e^{-2.52 \times 10^{-4} t_{burn}} \quad (\text{cooling}) \quad [3]$$

where ΔT was the daily post-fire temperature change ($^{\circ}\text{C}$), and t_{burn} was the time since burn, in days. The resulting temperature change was added to ECHAM5 A1b +T daily surface soil temperatures for the 10 years following the burn.

Relative effects of climate change vs. disturbance. I developed a relationship between time since fire and the change in surface soil temperature for boreal and arctic regions using measured temperature changes from plot-level field studies (Harden *et al.* 2006; Nakano *et al.* 2006; O'Neill *et al.* 2002; Viereck and Dyrness 1979) to tower studies that calculated temperature increases from changes in albedo (Chambers *et al.* 2005; Chambers and Chapin 2002; Liu *et al.* 2005). Increases in surface soil temperature post-fire were modeled daily using Equation 2. Temperatures increases due to burning persisted for 20 years but appeared negligible following sufficient vegetation regrowth (Wieder *et al.* 2009). The resulting temperature change was added to ECHAM5 daily

surface soil temperatures for the 20 years following the burn and was assumed to be zero after 20 years.

The 20c3 scenario was used as the control to evaluate the effects of disturbance alone and A1b +T +WT was used to evaluate the combined effects of climate change and disturbance at each of the sites.

CHAPTER IV

RESULTS

Daring Lake Fen: Model Calibration And Validation

I compared the modeled soil temperatures to observed soil temperatures at five depths. Generally, GIPL-peat fit the measured soil temperature profile at Daring Lake Fen well, capturing both the annual trends and timing of spring thaw (Fig. 2). Model fit improved with the addition of the multiplicative coefficients for thermal conductivity, C_f and C_t (Table 2). While the combination of $C_f=2.5$, $C_t=2.5$ did not yield the lowest RMSE values in the model validation, I used these parameter values because of a combination of low RMSE at most depths (Table 2), and accurately representing whether the soil was frozen or thawed in all years (Fig. 2). Alternately, increasing the soil moisture at the site achieved the same results, but model conditions do not then represent site conditions and the model is less easily transferred to other sites.

Climate Change Scenarios At Daring Lake Fen

Active layer depths between 1900-2000 ranged from 47.9 cm to 80.4 cm, with a mean of 63.3 cm. From 2001-2100, active layer depths ranged from 51.9 cm to 80.6 cm for the control (20c3) scenario, 51.2 cm

to 109.1 cm for A1b+T, and 53.9 cm to 108.9 cm for A1b+T+WT. Mean active layer depths were 67.0 cm, 76.6 cm, and 78.7, for control (20c3), A1b+T, and A1b+T+WT, respectively. Warmer air and surface soil temperatures and higher water table levels in the A1b scenario led to significantly deeper active layer depths by 2100 as compared to the control scenario (Table 5; Fig. 3). This was true for both warmer (A1b+T) and warmer and wetter (A1b+T+WT) scenarios. No significant difference was observed in active layer depths between the two A1b scenarios ($F(1,208) = 1.51, p=0.22$), indicating that the small differences in water table levels resulted in small differences in the active layer depths. Overall, temperature was the best predictor of maximum active layer depth across all scenarios ($B_0=0.9931, B_1=0.0425, F(3, 311)=313.3, p<0.0001$).

Mean active layer depths for A1b were an average of 25 cm deeper than active layer depths for 20c3 between 2090 and 2100. Similarly, the number of days with soil temperatures above freezing (thaw days) at 10 cm per year increased by 20% in the climate change scenario between 2090 and 2100, from 132.4 to 159.3 days for 20c3 and A1b +T +WT scenarios respectively (Fig. 4). Interestingly, the A1b+T from 2000-2010 show almost 10 fewer thawed days than the control run; this appears to be driven by higher water table levels between the two treatments (-7.9 cm in the control vs. -3.2 cm in the warming) rather than by differences in surface soil temperatures.

Temperatures at 60 cm, the bottom of the peat, increased in all climate scenarios, most notably for the A1b climate scenarios (Table 5). The rate of temperature increase at 60 cm and 2.5m was less than surface temperature increases, but still resulted in a temperature increase of 4.8°C at 60 cm and 4.5°C at 2.5 m in both A1b scenarios. Similarly, soils at greater depths were within the range where liquid water is found ($> -1^{\circ}\text{C}$) for a longer time in the climate warming scenarios (A1b) than the control scenario (Fig. 5).

Water Table Sensitivities

Maximum annual active layer depths using a water table ranged from 70.6 – 119.8 cm for 20% saturation, 60.1 – 110.8 cm for 60% saturation, and 54.6 – 107.8 cm for 100% saturation (water table at the surface). Significant differences in active layer depths resulted from using a water table with an unsaturated zone compared to using 20% or 60% saturation (Table 6; $F(3, 396)=63.93$, $p<2e-16$). Differences in active layer depths between using a water table and 100% saturation were not significant ($p=0.695$); differences between all other treatments were significant ($p<0.05$). Drier soils had thicker active layers.

I also explored the sensitivity of active layer depths to the seasonality of wetting and drying periods. Active layer depths were significantly shallower in a scenario where the water table was 10 cm lower during the growing season than in the A1b +T +WT scenario for 2001-

2100. Mean active layer depths for the 2001-2100 were 72.8 cm in the dry growing season scenario and 78.7 cm for the control scenario. Neither changing soil moisture during the spring and fall nor a trend in the water table significantly affected active layer depths (Table 7). As anticipated from the climate change scenario results, active layer depths for all scenarios were deeper from 2091-2100 than 2001-2010.

Organic Layer Sensitivities

The thickness of the organic layer was an important determinant in the depth of the active layer. In a simulation of the active layer at Daring Lake Fen from 1900 -2100, a peat thickness of less than 30 cm resulted in active layer depths that were approximately 100% deeper than active layers in simulations with 30 cm of peat or more (Fig. 6).

To further explore this relationship, I used differing peat thicknesses and both a dynamic water table and constant soil moisture at field capacity. I saw a logarithmic relationship between organic layer thickness and the mean active layer depth using the A1b+T+WT scenario (Fig. 7). The shape of the curve varied by soil moisture. Inclusion of a water table decreased active layer thickness relative to soils at field capacity, with the exception of the 15 cm organic soil. With the inclusion of water table, shallower peat depths had shallower active layer depths than the same peat depth with soil moisture field capacity. Additionally,

the annual rate of active layer increase was nearly twice as deep in shallow peat depths as in deeper peats (Table 6).

Disturbance Scenarios

In this section, I present results in two formats. The first is in relation to the current, post-burn peat surface. The second is in relation to the pre-burn peat surface elevation; that is the post-burn peat surface plus the burn depth.

Sensitivity analysis: Factors affecting post-burn active layer depths. I explored the differences between soil temperatures and active layer depths of soils that were burned early in the season (June 27) vs. late in the season (Sept. 23). The timing of the burn had no effect on annual active layer depths within the year of the fire or when all years were considered ($F(1, 103)=0.5650, p=0.4540$). There were no significant differences in soil temperatures at 10 cm, 60 cm or daily active layer depths between burns that happened in the early season (June 27) or late in the season (23 September), over the three years, five years and ten years after the simulated disturbance in the disturbance + climate change scenario.

Similarly, fire severity, as represented by the burn depth, did not significantly affect daily active layer depths within the first year ($F(2, 4014)=0.002, p=0.9875$) or over the next decade ($F(1, 35772)=0.3133,$

$p=0.5756$). Peat temperatures at 60 cm depth were not significantly different between burn depths ($F(1, 10945)=0.0215$, $p=0.8834$), but burn depth was an important predictor of maximum active layer depth in 2019, the burn year ($F(1,7)= 8.1183$, $p=0.02471$).

While active layer depth wasn't different between fire severity treatments, I saw significant differences after adjusting the annual active layer depths to the original pre-fire peat surface (i.e. adding the burn depth to the active layer depth). There were significant differences between active layer depths, in addition to differences between post-fire peat temperatures relative to the pre-burn surface elevation (Table 8).

Post-burn surface soil temperature response was a significant predictor of modeled differences in active layer depths and temperatures ($F(2, 42981)= 854.8$, $p<2e-16$). I found significant differences in daily and annual active layer depths between control (A1b+T+WT), burned and warm (A1b+T+WT+B), and burned and cool (A1b+T+WT-B) scenarios (Table 8). Mean active layer depths for the control treatment (no post-fire temperature change) were 67.6 cm, while mean active layer depths for warmer post-fire surface soil temperatures were 17.4 cm deeper and active layer depths for cooler post-fire surface soil temperatures were 11.3 cm shallower for the 9 years post-fire (Table 8).

Relative effects of climate change vs. disturbance. For the first twenty years after the simulated wildfire, post-fire annual active layer depths increased by 10 cm and 7 cm in the unburned control (20c3) and climate change scenarios (A1b +T +WT), respectively (Fig. 8, 9). Between 2040 and 2100, post-fire active layer depths were 1.6 cm deeper and 0.4 cm shallower than in the unburned control (20c3) and climate change scenarios (A1b +T +WT). By 2100, active layer depths in burned and unburned A1b + T +WT scenarios were 25 cm deeper than in the burned and unburned control (20c3) scenario.

Annual differences between the burned and control runs from 2040 to 2100 were correlated with differences in water table position (Fig. 10). The correlation was significant; F-values range from 10.87 to 127.51 with 1, 59 degrees of freedom and p-values range from <0.0001 to 0.0017. The R² of these relationships ranged from 0.16 to 0.68.

When the effects of wildfire were adjusted to consider the active layer depth relative to the original, pre-burn peat surface, the active layer depth was an additional 5 and 20 cm deeper in the burned scenarios than in the unburned scenarios. This resulted in significantly deeper active layers in burned scenarios than unburned scenarios (Table 9).

One other place where wildfire causes a small difference is in soil temperature at depth (Fig. 11). While mean annual soil temperatures do not differ between burned and unburned soils, the range of depths and

length of time when soil temperatures are between -1°C and $+1^{\circ}\text{C}$ is greater and longer in the burned A1b scenario than the unburned A1b scenario.

CHAPTER V

DISCUSSION

Relationships Between Organic Soil, Water Content, And Thermal Conductivity

Permafrost exists in disequilibrium with current climates due to the thermal offset and ecological protection in many areas. However, I found it difficult to capture annual permafrost dynamics and soil temperatures using published estimates of thermal conductivity for peat or organic soils during my validation runs. Modeled soil temperatures at 40 cm and 60 cm remained frozen during the growing season, while observed soil temperatures were above freezing at Daring Lake Fen (Fig. 2).

With the inclusion of multiplication factors (C_f and C_t) to increase the thermal conductivity of the fen peat for both frozen and thawed conditions, I observed very good correlation between the modeled and measured soil temperatures, although the model still underestimated the number of thaw days at these depths (Table 2). Using the correction factors, thermal conductivity values of bulk soil ranged from 1.19 to 4.12 $W m^{-1} K^{-1}$, generally higher than literature values for moss and peat soils with varying water contents that range from 0.02 to 0.61 $W m^{-1} K^{-1}$ for thawed

soils (c.f. O'Donnell *et al.* 2009; van Wijk and de Vries 1963) and 0.80 to 2.65 $\text{W m}^{-1} \text{K}^{-1}$ for frozen soils (Anisimov *et al.* 1997).

Alternatively, increasing the amount of soil moisture within the peat by increasing the water table level by ~7 cm produced similar results as the inclusion of the multiplication factor on thermal conductivity. A separate analysis confirmed that thermal conductivity was most sensitive to changes in water content; an order of magnitude increase to the thermal conductivity of peat (from $0.06 \text{ W m}^{-1} \text{K}^{-1}$ to $0.6 \text{ W m}^{-1} \text{K}^{-1}$) resulted in an $0.2 \text{ W m}^{-1} \text{K}^{-1}$ increase of the soil thermal conductivity, while increasing the water content from 60% saturation to 100% saturation resulted in an $0.8 \text{ W m}^{-1} \text{K}^{-1}$ increase of the soil thermal conductivity and a 4.5 – 6.8 cm shallower active layer depth (Table 6).

While the effects of soil moisture on soil thermal properties was large, changes in soil moisture seasonality do not seem to significantly affect active layer depths, with the exception of drier growing seasons/wetter non-growing seasons (Table 7). Wetting and drying trends in soil moisture do not produce a trend in active layer depths, but does seem to play a secondary role in determining annual differences in post-fire active layer depths following vegetation recovery (Fig. 10). Moisture doesn't seem to be the key driving factor in the model, however, as the response of permafrost to temperature changes is much larger than the

response to trends in soil moisture. Changes in soil moisture in a drier site might be more significant.

Relict permafrost often exists in disequilibrium with the climate in peatlands due to the combination of organic soil thickness and soil moisture (i.e. the low thermal conductivity of dry peat). Results from the sensitivity analysis indicate an interesting interaction between soil moisture and organic soil thickness (Fig. 7). No significant differences in mean active layer were found between 22 cm, 30 cm and 60 cm of peat, but active layer depths using 15 cm of peat were 30 cm to 60 cm deeper than thicker peat, or two-thirds to nearly twice as deep (Fig. 7). The difference between the permafrost response to shallower organic soils appears to be due to an interaction between organic soil thickness and water content (Table 6). This indicates that there may be either a soil moisture or peat depth threshold at which the other factor becomes more important in determining the active layer depth of the permafrost.

An interaction between organic soil thickness and soil moisture has implications for a non-linear response to changing climates. If increased decomposition occurs in deep peat following warmer soil temperatures (Dorrepaal *et al.* 2009) and leads to a either a decrease in the peat accumulation rate or the organic soil thickness without a change in water table, my results indicate a subsequent, non-linear change in the depth to the permafrost table (Fig. 7). This may result in talik formation and

ultimately lead to permafrost degradation, although this isn't predicted for Daring Lake Fen by 2100. Warmer soil temperatures also may lead to a positive feedback to decreasing peat thickness as respiration increases with warmer temperatures, resulting in warmer soils.

My predictions of permafrost stability in the future were likely somewhat conservative. First, ECHAM5 modeled winter temperatures were lower than measured air temperatures at Daring Lake Fen and resulted in colder soil temperatures. Had the ECHAM5 predictions matched my site data, I would have anticipated warmer surface soil temperatures and subsequent warmer temperatures throughout the profile, resulting in deeper active layers. Additionally, model validation showed an underestimation of thaw days at depth in the peat, indicating that the model was also biased towards lower soil temperatures. Both factors combine to lead to conservative estimates of permafrost persistence; in reality, soil temperatures in the future will likely be warmer than predicted by the model and ECHAM5 data.

Relative Effects Of Climate Change Versus Disturbance

Temperature increases predicted by ECHAM5 climate change scenarios resulted in significant increases in both active layer depths and soil temperatures at Daring Lake Fen by 2100. The largest temperature increase occurred in the winter, a +8.3°C air temperature increase.

However, mean annual temperatures at Daring Lake Fen remained below 0°C, and I did not simulate permafrost degradation or talik formation. It appears unlikely that substantial ecosystem-level changes or permafrost collapse will occur at this site, despite a 30-35% increase in active layer depths. Other simulations of permafrost stability in Canada show similar results and predict no degradation of permafrost at Daring Lake Fen (Zhang *et al.* 2008).

I did observe increases in both active layer depth and soil temperatures for the climate change scenarios. In the A1b climate change scenarios, the soil temperatures at 2.5 m are above -1°C for more of the year than the control scenarios (Fig. 5). When soil temperatures are above -1°C, there is more liquid water as phase change has begun to occur and thus increased rates of biological activity (Panikov *et al.* 2006). Similarly, the range of depths and length of time when soil temperatures are between -1°C and +1°C was greater and longer in the burned A1b scenario than the unburned A1b scenario (Fig. 11).

While *Kasischke and Johnstone* [2005] suggest that consideration of burn timing is important in determining ecosystem response to wildfire, I observe no differences between post-fire active layer depths in later season burns and earlier season burns (Table 8). Feedbacks between burn timing, burn severity, and active layer depth are not accurately represented within the model because the water table remains constant

throughout the year and burn depth is an input parameter. Because Daring Lake Fen refreezes completely at the end of the season, burn timing does not have a direct effect on active layer depths beyond the first year.

Of all hypothesized controls on post-fire active layer depths, significant differences were found only between post-fire surface soil temperature response treatments. Burn depth has very little effect on soil temperatures and active layer depth (Table 9), but it is important when considering the change in active layer depth relative to the pre-burn peat surface elevation, which is potentially important for biological processes. There may be indirect effects of greater burn depths that directly affect surface soil temperature, such as post-fire regeneration. Higher severity burns resulting in slow post-fire vegetation recovery (Benscoter *et al.* 2005; Johnstone and Chapin 2006) may lead to higher post-burn surface soil temperatures through low albedo from charred surfaces and from lack of shading. This will result in deeper active layer depths that are dependent on burn severity and post-fire recovery rather than burn depth directly.

My results indicate that the effects of fire on active layer depth of permafrost are relatively short-lived and essentially confined to the period of altered albedo and increased surface soil temperatures (Fig. 8, 9). I assumed post-fire temperature changes lasting 20-years due to

vegetation recovery, but albedo may actually increase to levels higher than unburned stands in less than 10 years post-fire (Randerson *et al.* 2006). If fire frequency increases to the point where surface temperatures do not recover to pre-fire temperatures, I would expect to see additional permafrost degradation due to higher surface soil temperatures. After the temperature returns to the control run (no burn temperature), the differences in active layer depths between burned and unburned runs after twenty years are generally less than 5 cm and are correlated with differences in water table levels (Fig. 10).

Changes in albedo, whether due to charred post-wildfire soils, vegetation recovery, or flooding, can have a significant effect on surface soil temperatures. The effect may be larger than I capture in my model because I consider only the changes in surface temperature due to charred surfaces. Changes in albedo occur in flooded or fully saturated soil; the albedo of open water is much lower than albedo of sphagnum moss and other peatland species and can result in a significant temperature increase at the surface (J. Harden, pers. comm., 2010; Harris 2002).

Implications For Soil Carbon And Biogeochemistry At Daring Lake Fen

Changes in the active layer depths and soil temperatures of permafrost sites have implications for the carbon balance at Daring Lake

Fen. With the climate change scenarios at Daring Lake Fen, I observed an increase in growing season length by nearly a month, a 30-35% increase in active layer depths, and warmer soils at depth. All of these changes in soil temperature lead to increases in the potential zone of and duration of biological activity (Fig. 5).

The increase in growing season length was derived from the number of days above freezing within the rooting zone (top 10 cm). A recent increase in growing season length (measured as the number of days between freezing and thawing of surface soils) in northern latitudes has been found by others (Euskirchen *et al.* 2006; Smith *et al.* 2004) and my results indicate this will continue into the future. Euskirchen, *et al.* [2006] found a rate of increase in the growing season of 0.38 days yr⁻¹ from 1960 -2000 and predicted a rate of 0.35 days yr⁻¹ for 2000 - 2100. They found that increases in growing season length were correlated with increases in net ecosystem productivity, and vegetation C. The soil C response to growing season was bi-modal; soil C decreased with changes in growing season length of less than 6 days per century. Soil C increased with more than 5 additional days in the growing season per century. I find a slightly lower predicted increase in growing season of 0.25 days yr⁻¹ than Euskirchen, *et al.* (2006), but the increase in growing season length between 2000-2009 and 2090-2099 was 25 days. Therefore, I would expect to see higher NEP, vegetation C and soil C at Daring Lake Fen due

to the increases in growing season length. This is supported by results from Daring Lake; *Lafleur and Humphreys (2007)* observed higher CO₂ uptake in a year with an early spring (snowmelt occurred 3 weeks earlier than other years) and warmer air and soil temperatures, as compared with the other years in the study.

Increases in active layer depth and soil temperatures have essentially the same effects as each other. I expect the temperature increases to strongly affect C accumulation and C processes that are especially sensitive to changes in temperatures deeper in the soil (*Carrasco et al. 2006*). Organic matter from deep in the soil profile was generally mineralized with temperature increases following permafrost thaw (*Schuur et al. 2009; Zimov et al. 2006*). In organic soils with thawing permafrost, C emissions appear to be from the decomposition of deep peat (*Dorrepaal et al. 2009*). Due to the relatively shallow peat depths at Daring Lake Fen (60 cm), both increases in active layer and the increase of liquid water from warmer soil temperatures occur in the mineral soil. Permafrost regions often have higher amounts of C stored in mineral soil than non-permafrost soils, which can be due in part to cryoturbation (*Schuur et al. 2008*). If warmer soil temperatures do result in increased C mineralization rates and also result in increased vegetation productivity, vegetation C storage, and soil C storage, C uptake will need to increase to compensate for increased mineralization losses.

There is also a potential for changing the main pathway of C emissions at Daring Lake Fen. Because the site is a peatland site with a water table within 10cm of the surface, it is likely that methane (CH₄) production will become more important as a pathway of C release. In other boreal peatlands, CH₄ emissions are very sensitive to changes in soil temperature, especially with high water table levels (Turetsky *et al.* 2008). This potential change in the pathway of C emissions with temperature increases may have implications for the radiative forcing of this site.

CHAPTER VI

CONCLUSIONS

Climate change will significantly increase growing season length at Daring Lake Fen. Temperature is the most important predictor of changes in active layer depth and permafrost persistence, rather than changes in soil moisture. While others have found that changes in snowpack alter permafrost dynamics (Zhang *et al.* 2008), I wasn't able to capture differences in active layer depths or temperatures due to altered snowpack timing and amount.

Due to the importance of temperature effects on permafrost active layer depth, I found that the response of permafrost to wildfire is relative short and limited to the period of post-fire temperature changes and vegetation recovery, which I set to twenty years. Therefore, I anticipate that climate change will have much larger effects than wildfire on permafrost stability at this cold site, although this could change with significant decreases of the organic layer thickness. Thinner organic soils could result from repeated fires or other disturbance, or a significant mineralization due drier and warmer conditions at the site.

Increases in air temperature due to climate change will affect surface soil temperature and also significantly increase the growing

season length and soil temperatures at depth. These changes may result in increased vegetation productivity, NEP, and soil C storage. Soil C respiration will likely also increase due both to warmer temperatures and more thawed substrate as active layer depths increase. The C release from this previously frozen soil will probably consist of older C, but the C loss may be offset by the increased productivity. Still, because mean annual temperatures are predicted to remain below freezing, I predict that permafrost at Daring Lake Fen will remain relatively stable, and the site may even become "greener" due to higher vegetation productivity and increased C uptake.

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Table 1. Parameter values for the GIPL model. The model uses 6 soil layers: 3 peat layers with different properties, a mineral soil layer, and two bedrock layers. The volumetric water content (VWC) in the peat layers is determined using the water table functions.

Soil layer:	Peat 1: Fibric	Peat 2: Mesic	Peat 3: Humic	Mineral soil	Bedrock 1/ Bedrock 2
Soil properties:					
Thickness (m)	0.05	0.10	0.45	9.4	20 / 70
Porosity	0.95	0.92	0.9	0.62	NA
soil composition	100% peat	100% peat	100% peat	95% mineral 5% organic	bedrock
Water table & water content:					
Unsaturated zone VWC				0.43	0.96 / 0.42
<i>Water table functions:</i>					
When WTD < 10 cm from node	1 - 9 * distance ^a	1 - 5* distance ^a	1 - 1.33* distance ^a		
When WTD > 10 cm from node	0.1 - distance * 0.1	0.5 - distance * 0.6	1 - distance * 1.33		
Minimum water content	0.03	0.03	0.03		
Soil thermal properties:					
	Peat	Water	Ice	air	Mineral soil
Heat capacity (J m ⁻³ K ⁻¹)	3.44×10 ⁴ * VWC*100 + 583333	4.18×10 ⁶	1.9×10 ⁶	1.25×10 ³	2.0×10 ⁶
Thermal conductivity (W m ⁻¹ K ⁻¹)	0.06	0.57	2.2	0.025	2.0

^a Distance is depth of water table below node. If node was below water table, the water content was 1.

Table 2. Evaluation of model calibration parameters; C_f , C_t are multipliers on the thermal conductivity of frozen and thawed soils (including all soil components), respectively. Thaw days are mean number of days with temperatures greater than freezing per year; reported thaw day values are modeled - observed. Observed thaw days at Daring Lake Fen were 113.7 and 76.0 days at 40 and 60 cm, respectively.

Model	A	B	C	D
C_f	2.5	1.0	1.0	2.5
C_t	2.5	1.0	2.5	1.0
Thaw days @ 40 cm	-37.4	-61.0	-31.0	-64.0
Thaw days @ 60 cm	-51.0	-76.0	-40.0	-76.0
RMSE of modeled temperatures				
20 cm	1.956	1.947	2.145	1.799
40 cm	2.236	2.689	3.084	2.010
60 cm	2.611	3.192	3.822	2.398

Table 3. Comparison of ECHAM5 modeled temperatures, precipitation, and snowpack to observations at Daring Lake Fen and nearby Lupin A climate station.

	Daring Lake fen		Lupin A		ECHAM 5 20c3		ECHAM5 A1b	
Dates	2006-2008	1971-2000	1971-2000	1971-2000	2000-2100	2001-2010	2090-2100	2090-2100
Mean precipitation (mm)	104 (1997-2005)	131	201	266	196	193		
Annual precip. (mm)	299	494	480	578				
Maximum snow depth (cm)	37	138.2 (snow fall)	100 ^a	124 ^a	125 ^a	156 ^a		
Snow first day/last day	245/153	277/143	276/132	275/140	277/145			
Mean temperatures (°C)								
Air	1 Jan. - 31 Dec.	-8.7	-11.1	-8.2	-7.6	-7.7	-2.3	
Soil ^b		-3.5	NA	-6.7	-6.3	-6.4	-1.0	
Air	Winter (D-J-F)	-26.8	-28.5	25.0	-24.6	-25.5	-17.2	
Soil		-14.4	NA	25.5	-11.5	-25.9	-17.6	
Air	Spring (M-A-M)	-12.3	-15.5	-10.1	-8.6	-9.1	-3.9	
Soil		-10.8	NA	-7.5	-3.9	-6.8	-2.1	
Air	Summer (J-J-A)	10.8	8.9	10.3	10.9	10.9	14.4	
Soil		11.6	NA	13.1	11.6	13.7	17.6	
Air	Fall (S-O-N)	-6.8	-9.2	-8.2	-8.2	-7.7	-2.7	
Soil		-3.6	NA	-7.4	-1.4	-6.9	-2.2	

^aAssumes constant snow density of 150 kg m⁻³

^bSoil temperatures are measured at surface (below snowpack) at Daring Lake Fen and at 3 cm depth for ECHAM5.

Table 4. Input data for the climate change scenarios

	Control/20c3	A1b +T	A1b +T +WT
Inputs			
Temperature	ECHAM5 20c3	ECHAM5 A1b	ECHAM5 A1b
Precipitation	ECHAM5 20c3	ECHAM5 20c3	ECHAM5 A1b
Water table	Equation 1	Equation 1	Equation 1
Unsaturated zone water content	Table 1	Table 1	Table 1
Start date	1900	2001	2001
End date	2100	2100	2100
Spinup (yrs)	10	0	0
Initial soil temperature	Spinup	20c3	20c3
Trend, °C yr⁻¹			
Air temperature	0.008	0.065	0.065
Soil temperature	0.009	0.064	0.064

Table 5. Changes observed for inputs and results from GIPL-peat climate scenarios in 2001- 2010 and 2091-2100 using ECHAM5 modeled data.

	Years	20c3	A1b +T	A1b +T + WT
Air temperature (°C)	2001 – 2010	-7.72	-7.73	-7.73
	2091 – 2100	-7.27	-2.27	-2.27
	Trend (°C yr ⁻¹)	0.008	0.065	0.065
Surface soil temperature (°C)	2001 – 2010	-6.74	-6.68	-6.68
	2091 – 2100	-6.69	-0.86	-0.86
	Trend (°C yr ⁻¹)	0.009	0.064	0.064
Water table level (cm)	2001 – 2010	-7.9	-7.9	-3.2
	2091 – 2100	-8.1	-8.1	-3.8
Active layer depth (cm)	2001 – 2010	69	68	69
	2091 – 2100	67	89	91
	Trend (mm yr ⁻¹)	0.37	2.43	2.54
Thaw days at 10 cm	2001 – 2010	136.7	127.3	126.1
	2091 – 2100	132.2	156.8	157.1
	Trend (days yr ⁻¹)	0.043	0.35	0.35
Mean annual temperature at 60 cm (bottom of peat)	2001 – 2010	-8.04	-8.14	-8.11
	2091 – 2100	-7.76	-3.28	-3.28
	Trend (°C yr ⁻¹)	0.007	0.057	0.057
Mean annual temperature at 2.5 m (in mineral soil)	2001 – 2010	-8.03	-8.13	-8.12
	2091 – 2100	-7.81	-3.65	-3.69
	Trend (°C yr ⁻¹)	0.006	0.052	0.052

Table 6. Mean active layer and annual changes in active layer depths for sensitivity analyses. WT sensitivity is sensitivity to the distribution of water throughout the peat column; OL sensitivity is sensitivity to organic layer thickness. Water content refers to the method used to determine the water content above and below the water table (if applicable). Climate scenario A1b +T +WT was used for all sensitivity runs. The slope represents the annual changes in active layer depths (mm/yr), the intercept represents extrapolated active layer depths in 1900. All relationships are significant ($F(1,98) > 61.7$, $p < 0.0001$). There is a significant difference in the slopes between organic soil thicknesses (organic layer depth \times year interaction; $t = -3.324$, $p = 0.00094$).

Analysis	Organic soil thickness (cm)	Water content method	Water content in top 2m (m)	Active Layer Depths (cm)			Annual changes in active layer depths between 2001 - 2100		R ²
				2001 - 2010	2090 - 2100	Slope +/- error (mm/yr)	Intercept +/- error (cm)		
Climate change	60	WTD ^a	1.13	70	91	2.54 ± 0.30	40.4 ± 4.7	0.42	
WT sensitivity	60	20% saturation	0.40	85	106	2.36 ± 0.30	57.4 ± 4.6	0.39	
	60	60% saturation	0.69	76	97	2.45 ± 0.30	46.9 ± 4.7	0.40	
	60	100% saturation	1.15	69	91	2.51 ± 0.30	39.4 ± 4.6	0.41	
OL sensitivity	1	60% saturation	0.53	143	184	4.74 ± 0.57	87.8 ± 8.7	0.41	
	5	60% saturation	0.53	118	160	4.82 ± 0.56	62.1 ± 8.6	0.43	
	15	60% saturation	0.56	97	129	4.83 ± 0.55	40.6 ± 8.4	0.44	
	22	60% saturation	0.59	75	96	2.48 ± 0.30	45.9 ± 4.6	0.41	
	30	60% saturation	0.60	75	96	2.44 ± 0.30	46.6 ± 4.6	0.40	

Analysis	Organic soil thickness (cm)	Water content method	Water content in top 2m (m)	2001 – 2010	2090 – 2100	Slope +/- error (mm/yr)	Intercept +/- error (cm)	R ²
OL sensitivity	15	WTD	0.93	123	172	5.62 ± 0.64	58.0 ± 9.8	0.44
	22	WTD	0.97	71	91	2.46 ± 0.30	41.4 ± 4.6	0.41
	30	WTD	1.00	69	92	2.59 ± 0.31	39.6 ± 4.7	0.42
	60	WTD	1.13	70	91	2.54 ± 0.30	40.4 ± 4.7	0.42

a WTD: water content above the water table is determined through functions in Table 1. Below the water table, the peat is 100% saturated

Table 7. Timing of wet and dry periods during the year: mean active layer by decade. Climate scenario A1b +T +WT was used for all sensitivity runs. Active layer depths were significantly different between treatments when all years were considered ($F(6, 700)=3.90, p=0.0008$). Letters indicate significant differences between treatments. Differences between active layer depths for different treatments were not significant in 2001-2010 and 2091-2100.

A1b + T + WT			Active Layer Depths (cm)		
			2001 – 2100*	2001 – 2010	2091 – 2100
Control	Timing	A1b +T +WT	79 ^a	70	91
Drying trend	Annually	1 mm drier yr ⁻¹	78 ^a	69	92
Wetting trend	Annually	1 mm wetter yr ⁻¹	76 ^a	68	87
Wet	Spring	Dry fall	78 ^a	70	92
Dry	Spring	Wet fall	76 ^a	67	91
Wet	Growing season	Drier non-growing season	79 ^a	71	92
Dry	Growing season	Wetter non-growing season	73 ^b	63	87

Table 8. Burn sensitivity scenarios: mean active layer depths (ALD) from 2020-2028, relative to the new post-fire peat surface and to the original pre-burn surface elevation. All runs used the A1b +T +WT climate change scenario; burns occurred in 2019. Daily post-fire temperature change for the warmer scenario used Equation 2, while the daily temperature for cooler scenario change used Equation 3. Early season burn occurs on day 178 (27 June); late season burn occurs on day 266 (23 Sept.). Active layer depths were significantly different between post-fire temperature responses ($F(2,5) = 820.4, p < 0.0001$).

Depth burned (cm)	Post-fire temp. change	Burn timing	ALD, new post-fire surface (cm)	ALD, pre-burn elevation (cm)
0	None		68	68
5	None	Late	68	73
20	None	Late	68	88
5	Warmer	Early	83	88
5	Warmer	Late	82	87
20	Warmer	Early	82	102
20	Warmer	Late	82	102
5	Cooler	Early	57	62
5	Cooler	Late	56	62
20	Cooler	Early	56	76
20	Cooler	Late	56	76

Table 9. Post-fire active layer depths and soil temperatures for climate change plus disturbance scenarios. Post-burn surface soil temperatures were increased for 20 years using equation 4; post-burn water table was closer to the surface by the burn depth. For more details see Section 3.3.2.

Run	Depth burned (cm)	Burn Year	ALD, new post-fire surface (cm)		ALD, pre-burn elevation (cm)		Soil temps @ 2.5 m	
			2020-2029	2090-2100	2020-2029	2090-2100	2020-2029	2090-2100
20c3	0		65	67	65	67	-8.27	-7.81
20c3	5	2019	80	70	85	75	-6.57	-7.79
20c3	20	2019	79	70	99	90	-6.53	-7.73
20c3	5 + 20	2019 + 2080	80	72	85	97	-6.57	-7.23
Alb + I	0		68	89	68	89	-7.56	-3.65
Alb + I + WT	0		69	92	69	92	-7.59	-3.69
Alb + I + WT	5	2019	81	91	86	96	-5.89	-3.67
Alb + I + WT	20	2019	81	91	101	111	-5.86	3.62
Alb + I + WT	5 + 20	2019 + 2080	81	95	86	120	-5.89	-3.63

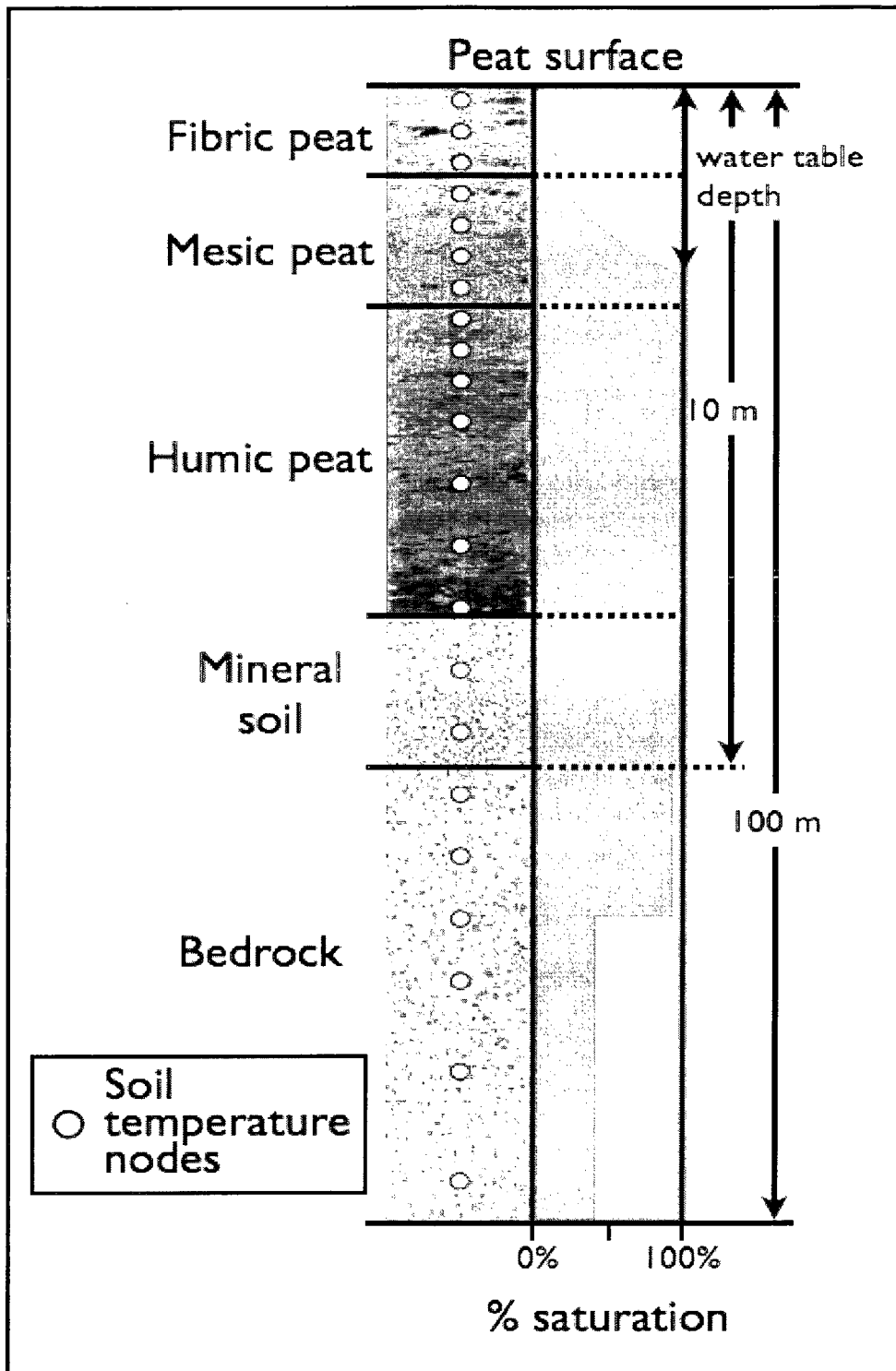


Figure 1. GIPL model schematic depicting the soil column including three peat layers, mineral soil, and bedrock; water table and percent of water filled pore space, and temperature calculation nodes used in the model.

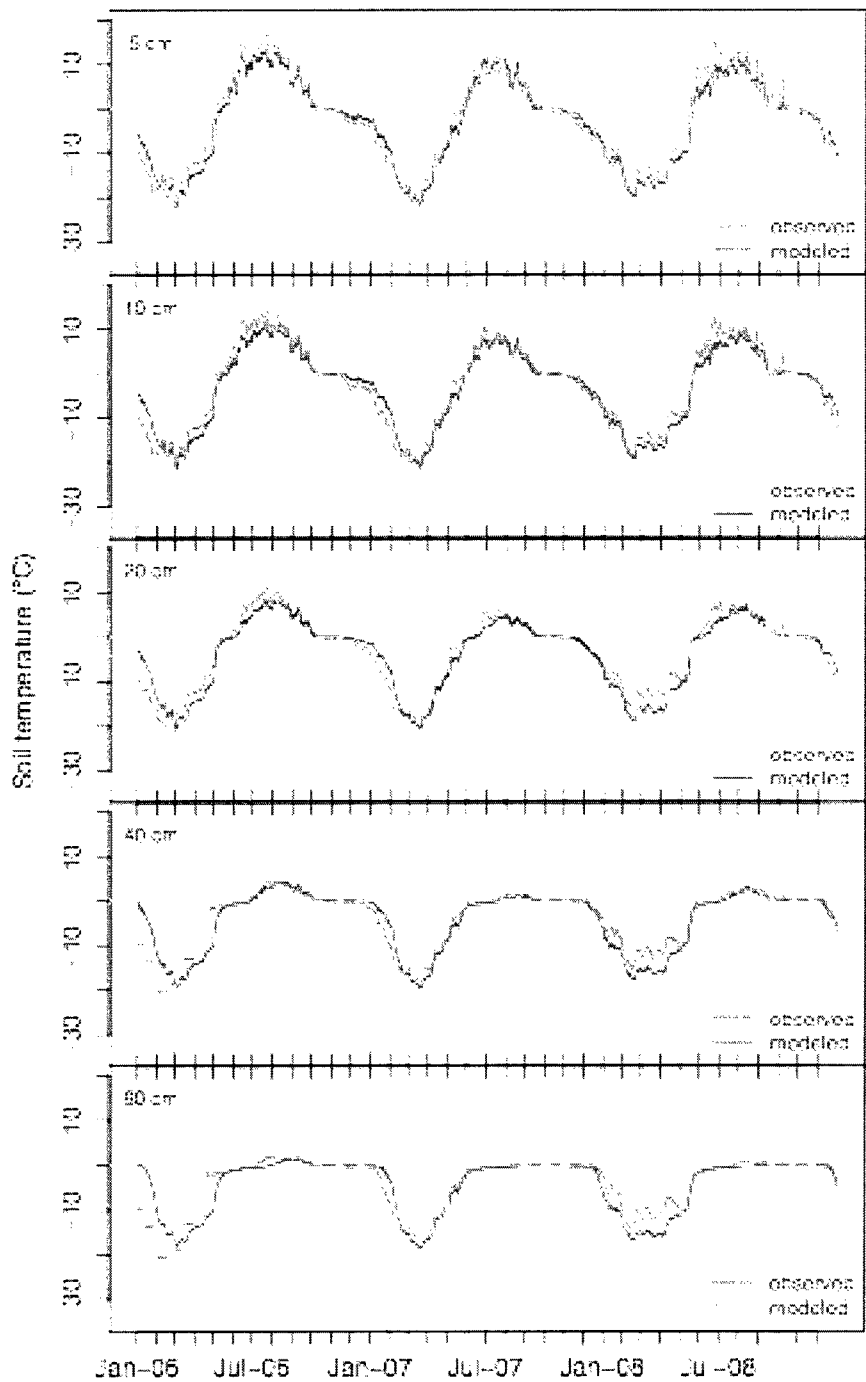


Figure 2. Observed and modeled soil temperatures at Daring Lake Fen from 2006 – 2008 using validation model A, with coefficients $C_f=2.5$, $C_r=2.5$ (Table 2).

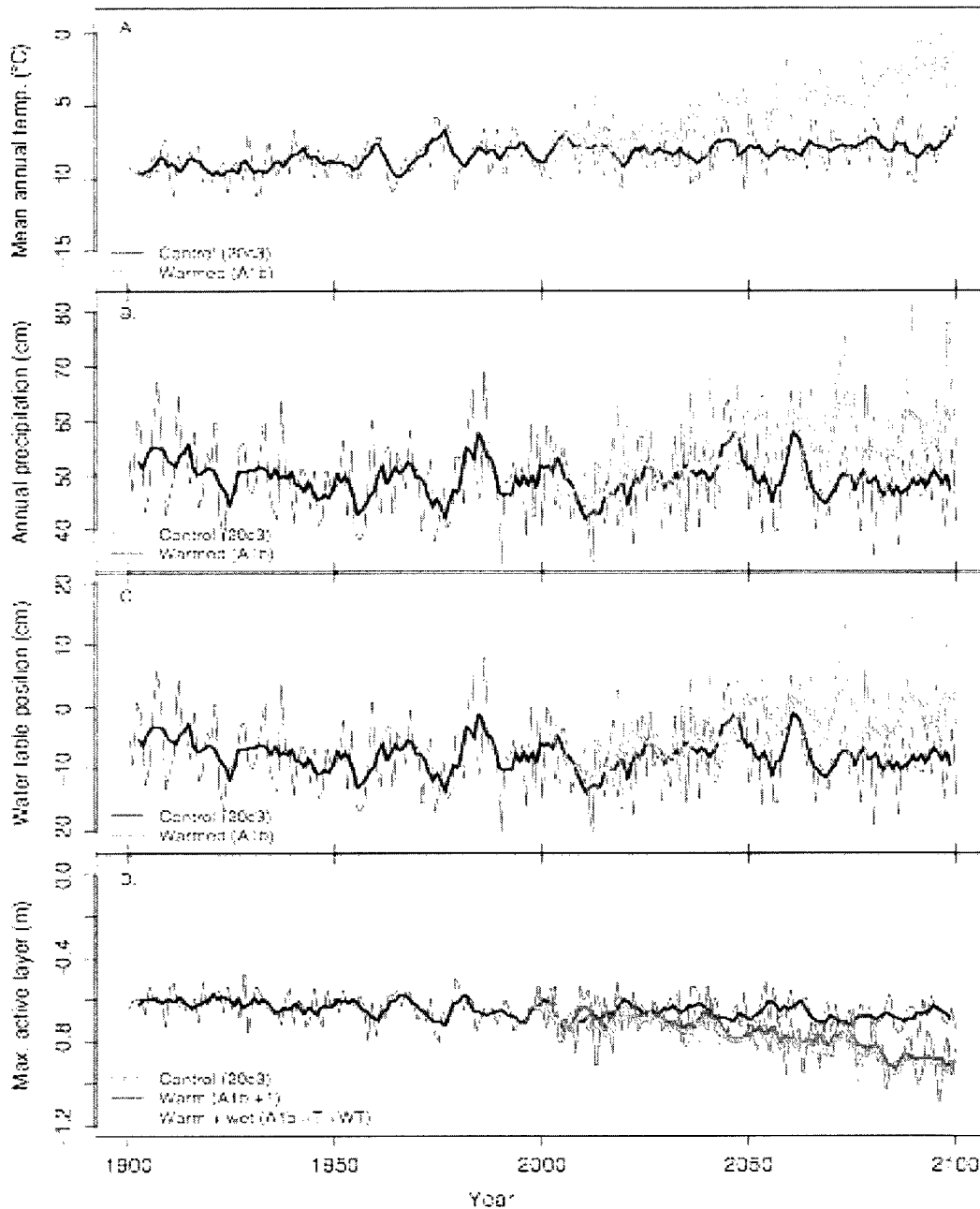


Figure 3. ECHAM5 modeled **(a)** mean annual air temperature, **(b)** precipitation and **(c)** mean annual water table position (negative values are below peat surface) at Daring Lake Fen from 1900- 2100 for control (20c3, black) and warmed (A1b, red) scenarios; **(d)** active layer depths at Daring Lake Fen for control (20c3, black lines) and climate change with (A1b +T +WT) and without (A1b +T) increased soil moisture (red and blue lines, respectively) scenarios. Bold lines indicate 5-year moving means. For further description of model scenarios, see Table 4 and Section 3.1.

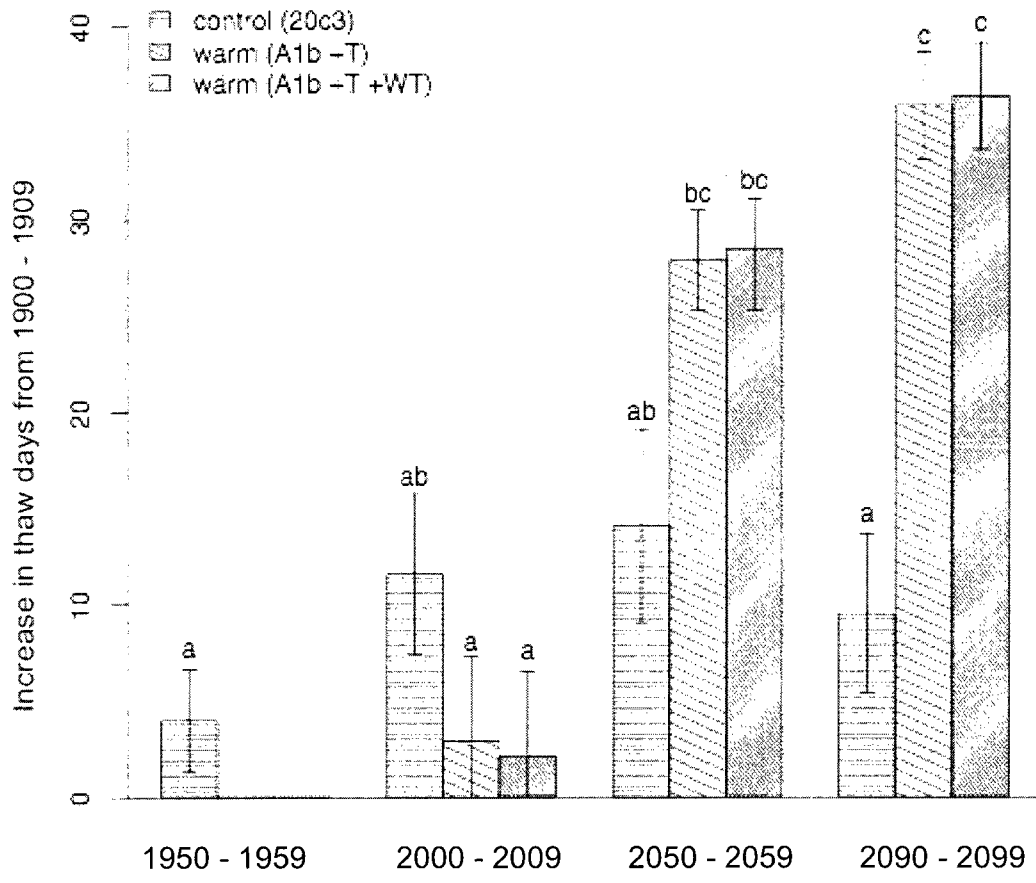


Figure 4. Increase in decade mean number of thaw days at 10 cm relative to number of thaw days in 1900-1909. Thaw days are days with the temperature above freezing. Letters indicate significant differences between decades and treatments.

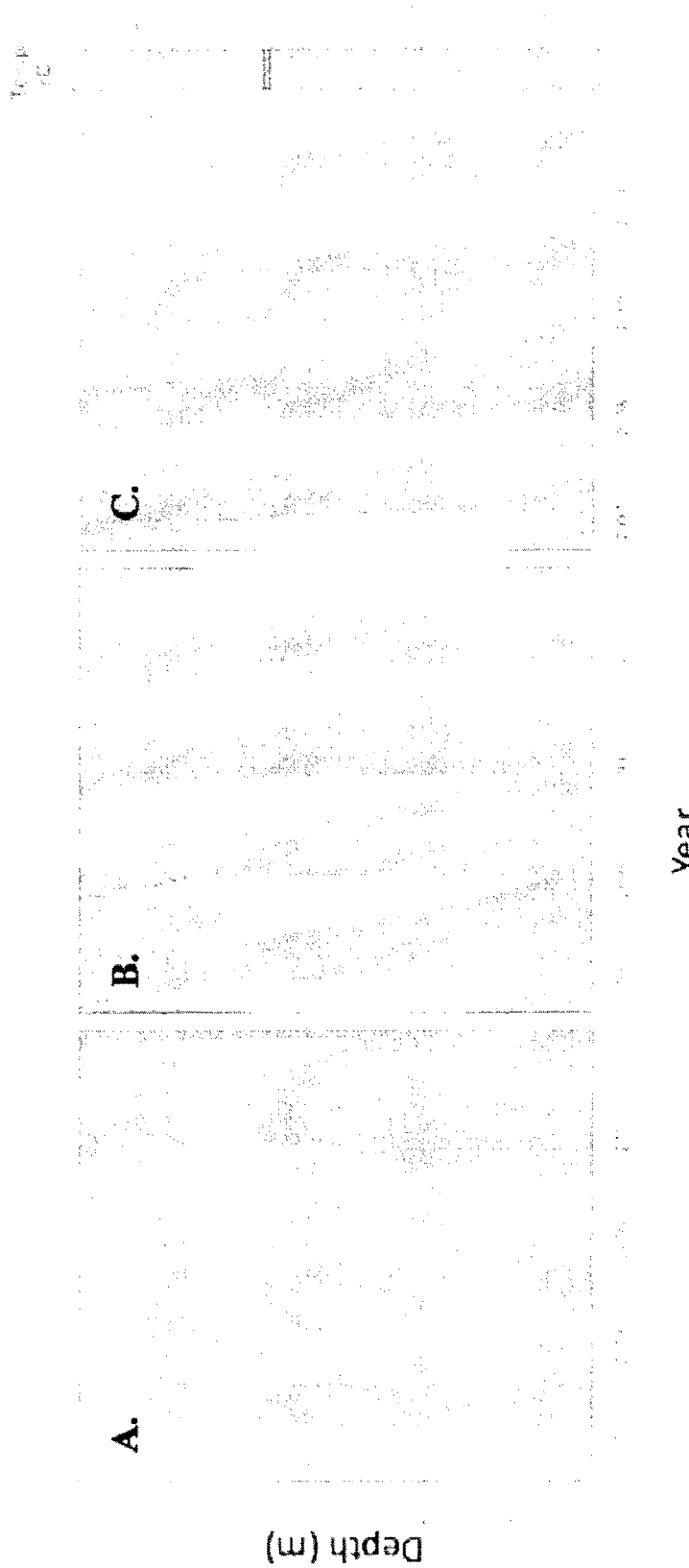


Figure 5. Soil temperatures from 0 to 3m depth in 2097-2100 for **(a)** control/20c3, **(b)** A1b +T, and **(c)** A1b +T +WT climate scenarios. Red indicates frozen soil (temperatures < -1°C), gold indicates soil temperature between -1°C and -0.1°C, green indicates soils where water is undergoing phase change (freezing/thawing), blue indicates soil temperatures between 0.1°C and 1°C, and purple indicates thawed soil (temperatures > 1°C). Soil temperatures from 2097-2100 range from -44.9°C to 29.6°C for control/20c3, and -33.2°C to 31.7°C for A1b +T and A1b +T +WT climate scenarios.

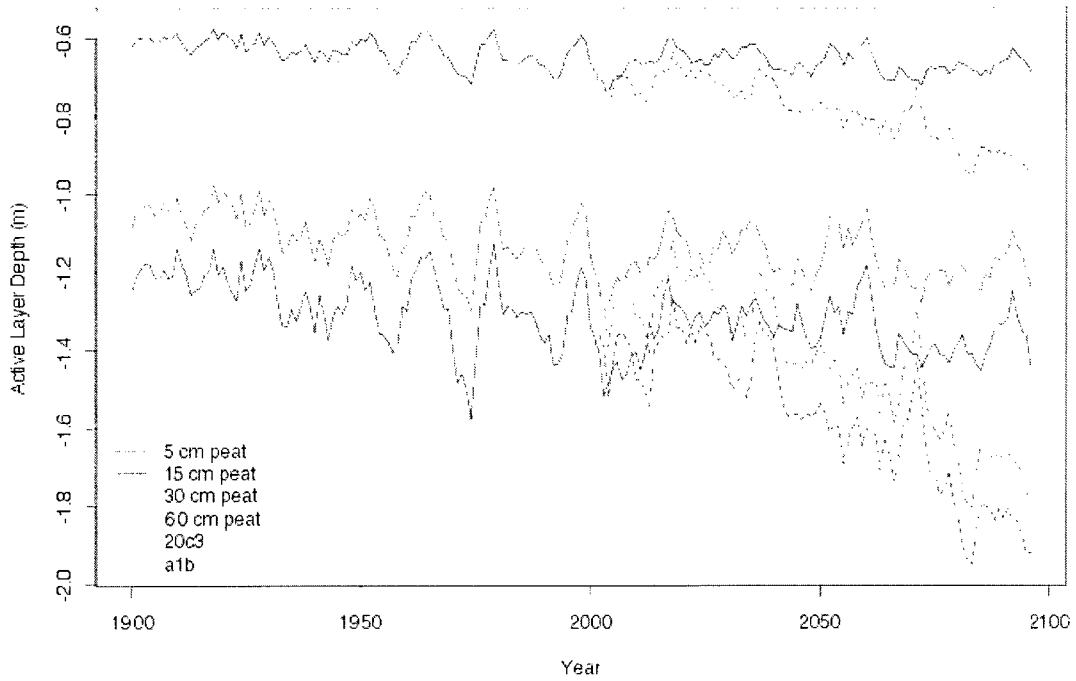


Figure 6. Five-year running mean of active layer depths using differing organic soil thicknesses for control (20c3) and climate change (A1b +T +WT) scenarios. Shallower organic soils (5 cm, 15 cm) are typical of upland soils, while thicker organic soils are typical of more poorly drained areas and peatlands. Water content in the unsaturated zone was a function of height above the water table for all peat depths.

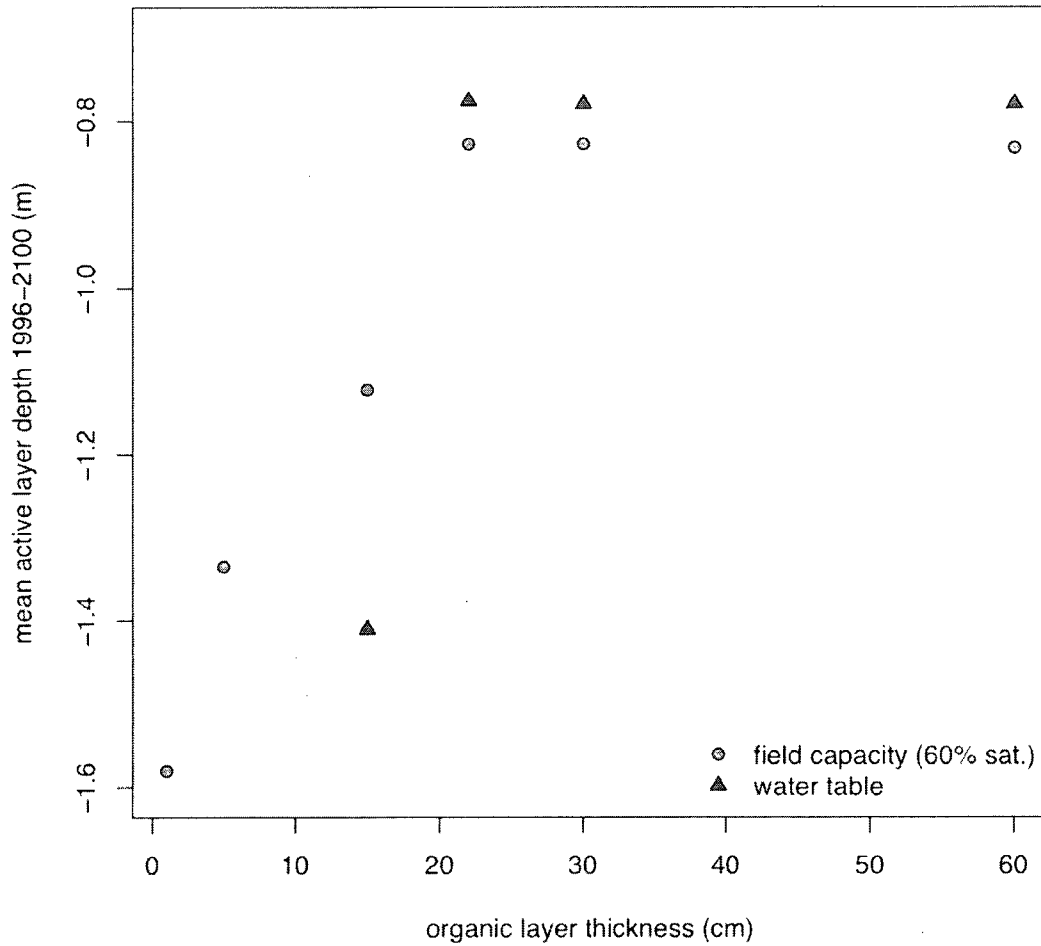


Figure 7. Relationship between peat thickness and mean active layer depth from 1996-2100 using the A1b +T (+WT) scenario. Symbols represent the method of determining the water content in the peat profile. Field capacity (red circles) assumes 60% saturation throughout soil profile. Water table (blue triangles) assumes the water content above the water table is determined through functions in Table 1, and below the water table, the peat is 100% saturated.

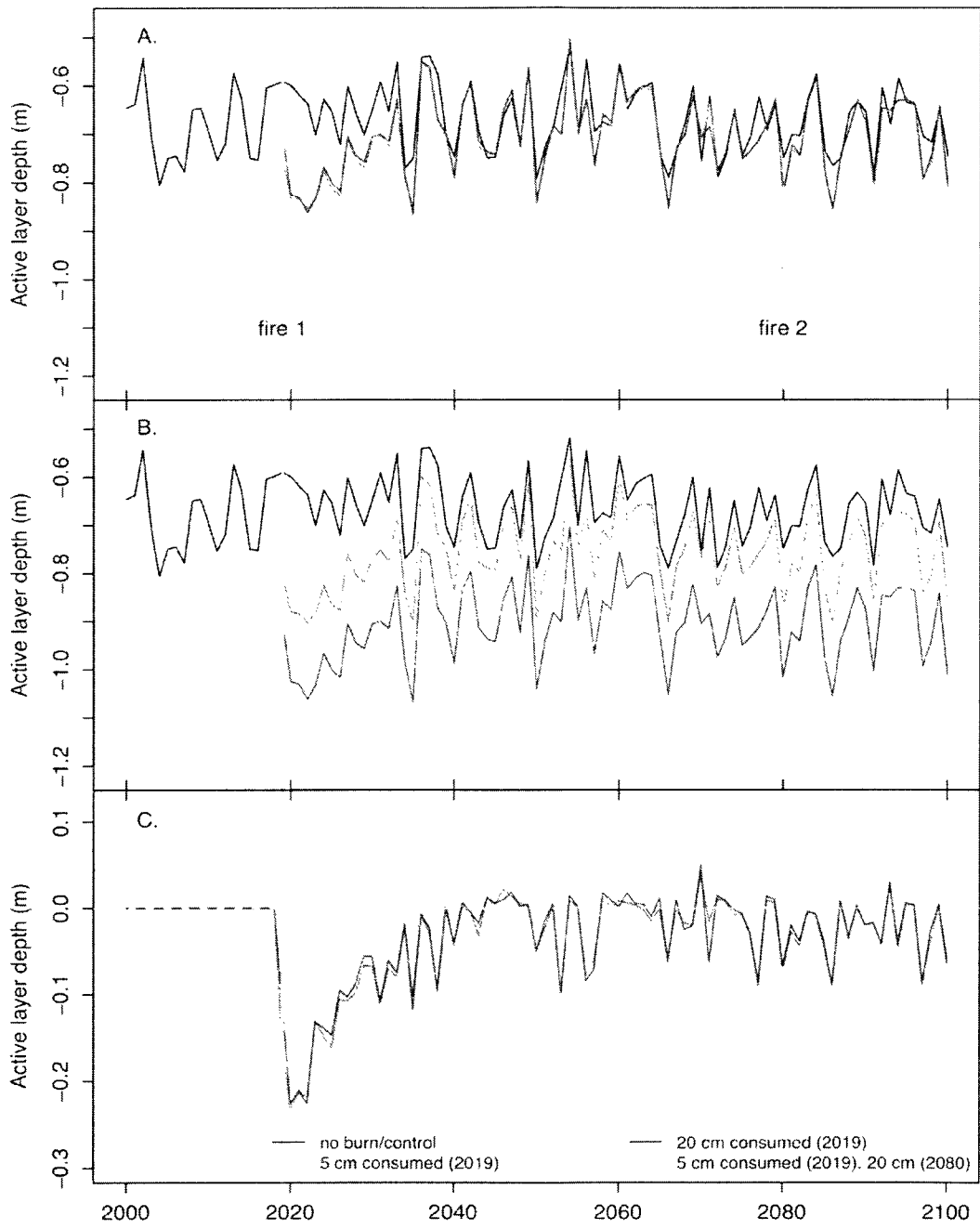


Figure 8. Post-fire active layer depths for control (20c3) scenario **(a)** relative to current peat surface, **(b)** relative to pre-burn peat surface, and **(c)** difference between burned and unburned active layer depths relative to current peat surface (not pre-burn surface).

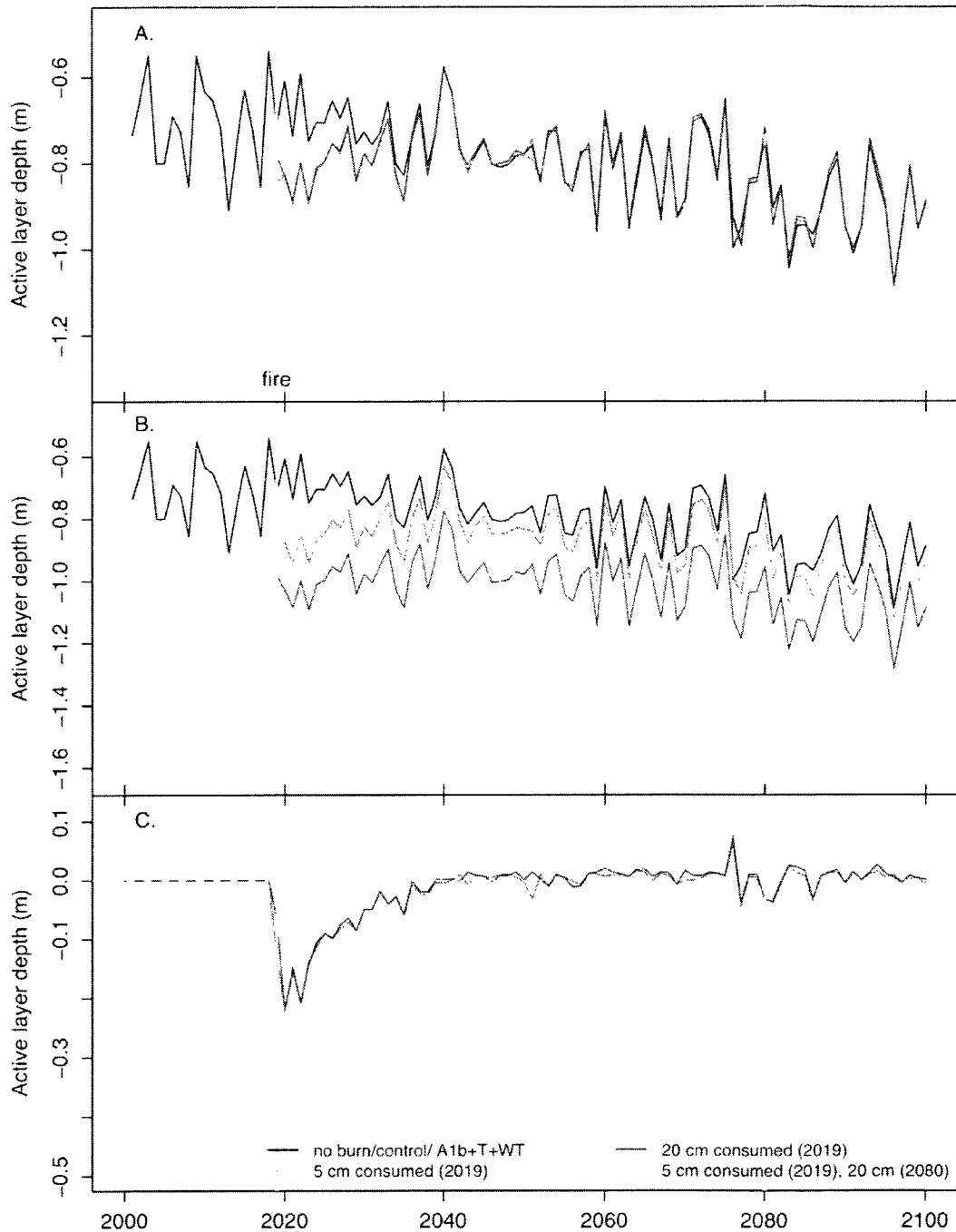


Figure 9. Post-fire active layer depths using climate change (a1b +T +WT) scenario **(a)** relative to current peat surface, **(b)** relative to pre-burn peat surface, and **(c)** difference between burned and unburned active layer depths relative to current peat surface (not pre-burns surface).

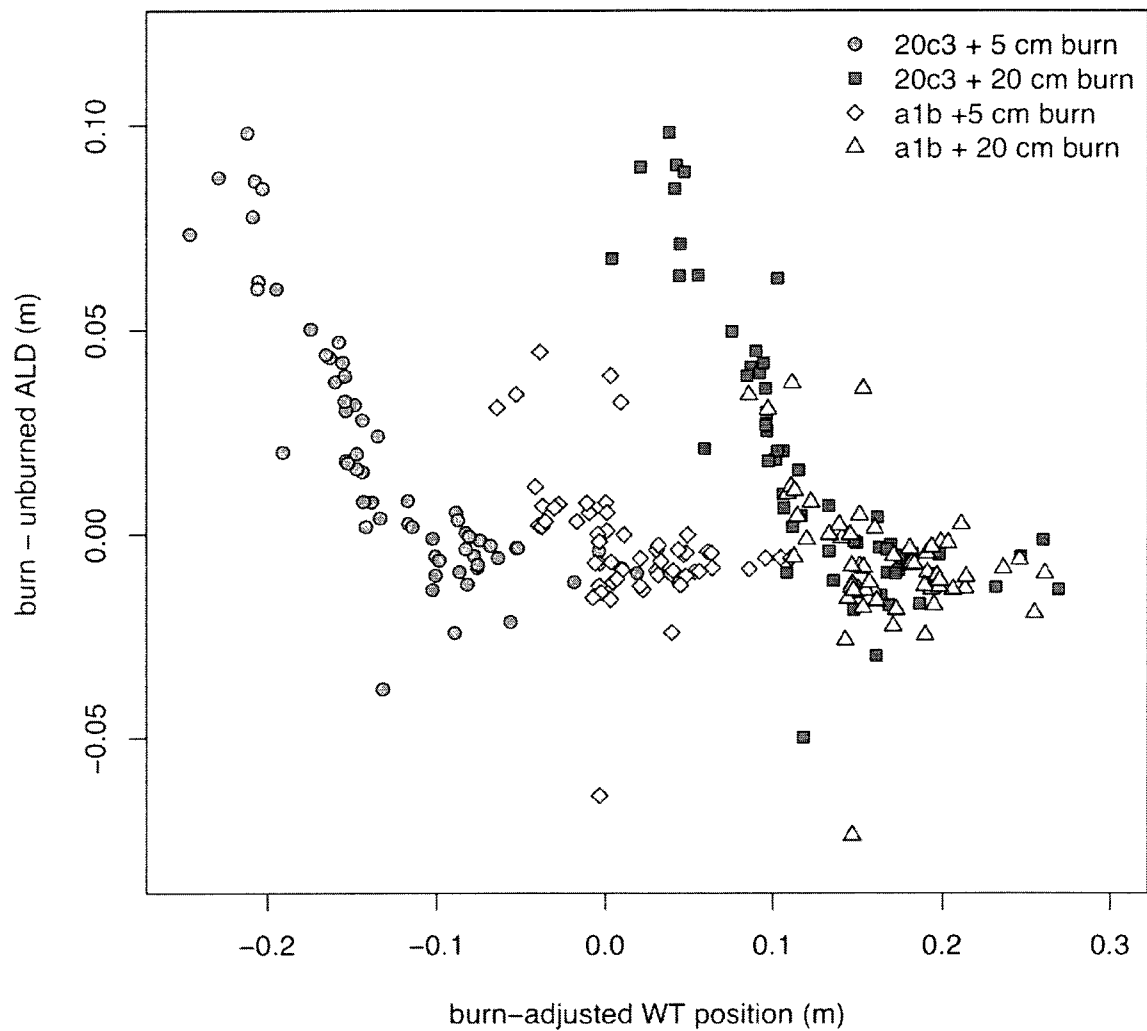


Figure 10. Differences in active layer depths in burned and unburned scenarios as a function of adjusted water table position (water table position + burn depth).

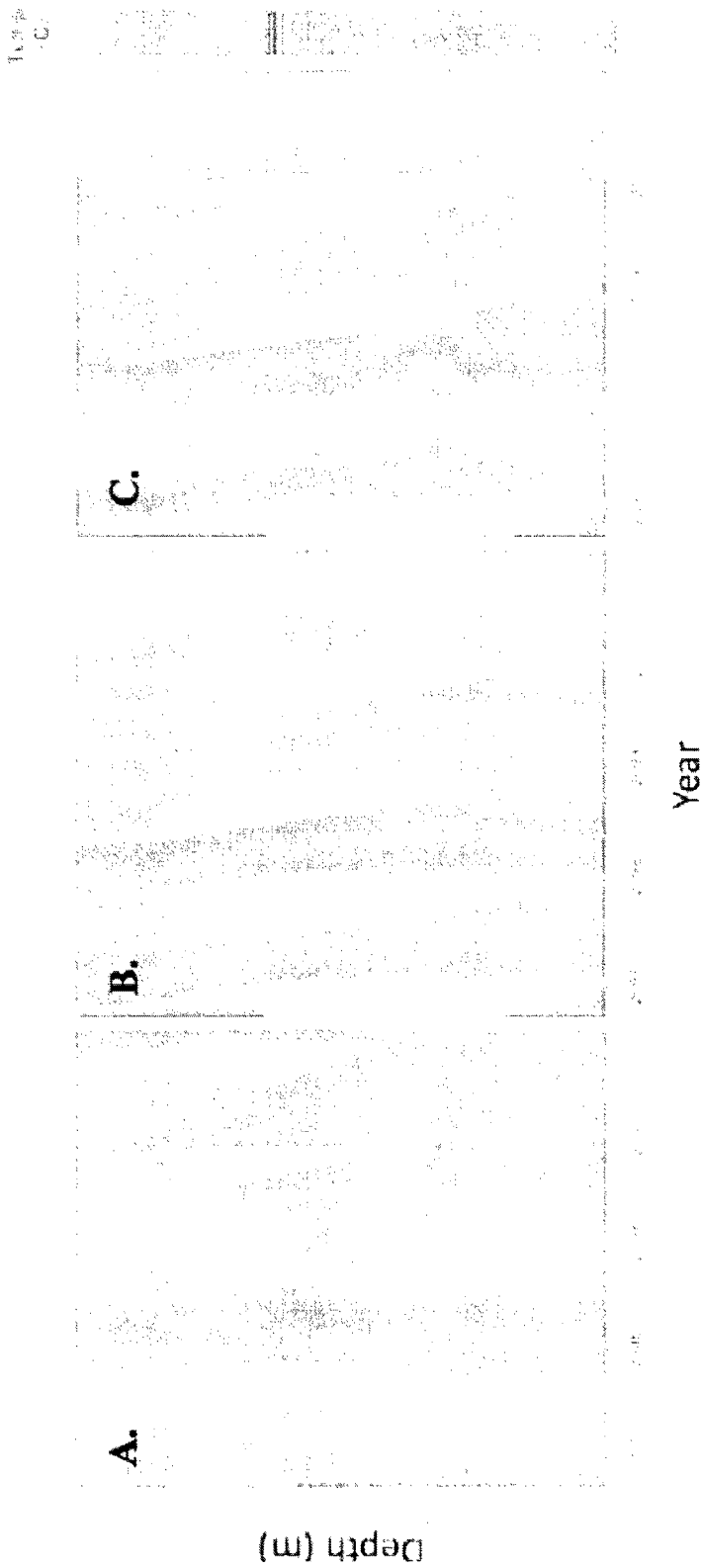


Figure 11. Soil temperatures from 0 to 3m depth in 2097-2100 for **(a)** A1b +T +WT control (no burn), **(b)** 5 cm burn, and **(c)** 20 cm burn scenarios. Red indicates frozen soil (temperatures < -1°C), gold indicates soil temperature between -1°C and -0.1°C, green indicates soils where water is undergoing phase change (freezing/thawing), blue indicates soil temperatures between 0.1°C and 1°C, and purple indicates thawed soil (temperatures > 1°C). Soil temperatures range from -33.2°C to 31.7°C for all scenarios.

APPENDIX

MODEL CODE FOR GIPL 2.0-PEAT IN MATLAB

```

airTemp= airTemp;           %[INPUT]
burnDay= burnDay;          %[INPUT]
runLength= length(airTemp);

% //Initial Temperatures and Depth
numSoilLayers=6;           %[6 SOIL LAYERS, TABLE 1]
TStep =24;                 %[TIME STEP, HOURS]
soilLayerDepth=[0.00 0.05 0.10 0.15 0.20 0.25 0.30 0.35 0.40
0.45 0.50 0.55 0.60 0.65 0.70 0.75 0.80 0.85 0.90 0.95
1.00 1.10 1.20 1.30 1.40 1.50 1.60 1.70 1.80 1.90 2.00
2.10 2.20 2.30 2.40 2.50 2.60 2.70 2.80 2.90 3.00 3.20
3.40 3.60 3.80 4.00 4.20 4.40 4.60 4.80 5.00 5.20 5.40
5.60 5.80 6.00 6.50 7.00 7.50 8.00 8.50 9.00 9.50 10.00
10.50 11.00 11.50 12.00 12.50 13.00 13.50 14.00 14.50 15.00 15.50
16.00 16.50 17.00 17.50 18.00 18.50 19.00 19.50 20.00 21.00 22.00
23.00 24.00 25.00 26.00 27.00 28.00 29.00 30.00 32.00 34.00 36.00
38.00 40.00 42.00 44.00 46.00 48.00 50.00 55.00 60.00 65.00 70.00
75.00 80.00 85.00 90.00 95.00 100.00]; %[NODE DEPTHS, METERS]
NumberOfSoilComputationNodes=114; %[# COMPUTATION NODES]
prevSoilTemp=zeros(1, NumberOfSoilComputationNodes);
bdepth= find(soilLayerDepth==depthBurned); %[LAYER OF BURN DEPTH]
D_init = soilLayerDepth; %[0 0.5 1 2 3 4 5 10 15 20 40 60 80
100]; %[DEPTHS, INITIAL CONDITIONS]
Dn_init = 14; %[# OF INITIAL DEPTHS]

%[CONVERGENCE CRITERIA]
E0= 0.0000014;
G0=0.015;
iter0=21;
maxABS = 1.41e-6;

%[UNFROZEN WATER FUNCTION]
a=[0.05 0.04 0.035 0.061 0.018 0.064];
b=[-0.1 -0.15 -0.32 -0.35 -0.17 -0.34];
c=[-0.1 -0.15 -0.32 -0.35 -0.17 -0.34];
ALFA0 = 20.14;
Tfr = 0; % evalin('base', 'Tfr'); % Temperature of freezing

```

```

FIT = 0.08; %evalin('base', 'FIT'); % "corridor of phase change, 0 degrees
+/- FIT. FIT = 0.08 degrees

```

```

% [INITIALIZE OUTPUT FOR SOIL TEMPERATURE, ACTIVE LAYER, THAW LAYER]
soilTempOut = zeros(NumberOfSoilComputationNodes, runLength);
activeLayerOut= zeros(1, runLength);
thawLayerOut = zeros(1, runLength);

```

```

% [SOIL PROPERTIES: SEE TABLE 1]
soilLayerThickness=[Zfib, Zmes, Zhum, (10 - Zfib-Zmes -Zhum), 20, 70];
p3 = (0.5*0.47 + (10 - Zfib-Zmes -Zhum-0.5 - 0.4)*0.42 + .4 * 0.8)/(10 - Zfib-
Zmes -Zhum); % calculate porosity in layer 4, variable thickness. Properties
for 0.5 m clay, rest coarse, from Wania Table 1.
porosity= [0.95 0.92 0.9 p3 1 1]; % 1 & 2 From Granberg 1999, original Sergei
stuff is used; null values
Speat = [1- porosity(1), 1- porosity(2), 1- porosity(3), Pmin, 0, 0] ; % volume
fraction of peat.
Smin = [0 0 0 (1-p3-Speat(4)) 0 0];

```

```

% [SOIL WATER CONTENT]
% Find volume fractions of each component in each layer, then the
volume within each layer (which is essentially the depth) & find water
content (% of pore space with water or % saturation) at each node for
top 2 meters
waterLayerDepth = [0.025 0.075 0.125 0.175 0.225 0.275 0.325 0.375
0.425 0.475 0.525 0.575 0.625 0.675 0.725 0.775 0.825 0.875
0.925 0.975 1.025 1.05 1.15 1.25 1.35 1.45 1.55 1.65
1.75 1.85 1.95 2 10 30 100];
waterLayerThickness = [0.025 0.05 0.05 0.05 0.05 0.05 0.05 0.05 0.05 0.05
0.05 0.05 0.05 0.05 0.05 0.05 0.05 0.05 0.05 0.05
0.05 0.05 0.05 0.025 0.1 0.1 0.1 0.1 0.1 0.1
0.1 0.1 0.1 0.05 8 20 70];
numberOfWaterLayers= length(waterLayerDepth);
Wsat= zeros(runLength, length(waterLayerDepth));
fmin = zeros(runLength, length(waterLayerDepth));
fpeat = zeros(runLength, length(waterLayerDepth));
fair = zeros(runLength, length(waterLayerDepth));
fwater = zeros(runLength, length(waterLayerDepth));
Z=Zfib + Zmes + Zhum;

```

```

%%%%% Soil moisture
if WTD == -1
    for i=1:1:length(waterLayerDepth-2)
        fpeat(:,i) = Speat(soilNumLayers(waterLayerDepth(i)));
        fmin(:,i) = Smin(soilNumLayers(waterLayerDepth(i)));
    end
    fwater= mfwater;
    fair = 1-fpeat - fmin - fwater;
%%%%%% Water table
else
    for j=1:1:runLength
        for i=1:1:length(waterLayerDepth-2)
            if soilLayerDepth(i) < WTD(j) && soilLayerDepth(i) < Zfib;
                Wsat(j,i)= nearSurfWat(soilLayerDepth(i), WTD(j));
            elseif soilLayerDepth(i) < WTD(j) && soilLayerDepth(i) < Zmes &&
                Wsat(j,i)==0;
                Wsat(j, i)= mesWat(soilLayerDepth(i), WTD(j));
            elseif soilLayerDepth(i) < WTD(j) && soilLayerDepth(i) < Zhum &&
                Wsat(j,i)==0;
                Wsat(j,i)= deepWat(soilLayerDepth(i), WTD(j));
            else Wsat(j,i)=1;
        end
    end
end
for i = 1: length(waterLayerDepth)
    fpeat(:,i) = Speat(soilNumLayers(waterLayerDepth(i)));
    fmin(:,i) = Smin(soilNumLayers(waterLayerDepth(i)));
    fair(:,i) = (1- Wsat(:,i)) * porosity(soilNumLayers(waterLayerDepth(i)));
    fwater(:,i) = Wsat(:,i) * porosity(soilNumLayers(waterLayerDepth(i)));
end
end

fwater(:,33) = 0.71*p3;
fwater(:,34) = 0.96;
fwater(:,35) = 0.42;
Wvol=fwater;

```



```

%[SOIL MOISTURE, BURNED SCENARIOS]
%The Burning Scenario (this isn't going to work if a site loses > 2m of
%soil. Currently, the organic soil does not re-grow post-fire.
if dayOfBurn > 1
    waterLayerDepthBurn = waterLayerDepth + depthBurned;
    burnSoilDepth=soilLayerDepth + depthBurned;
    for j=burnDay:1:runLength
        for i=1:(length(waterLayerDepth)-bdepth)
            fpeat(j,i) = Speat(soilNumLayers(waterLayerDepthBurn(i)));
            fmin(j,i) = Smin(soilNumLayers(waterLayerDepthBurn(i)));
            fwater(j,i)= fwater(j, i+ bdepth-1);
            fair(j,i) = 1-fpeat(j,i) - fmin(j,i) - fwater(j,i);
        end
    end
    fwater(burnDay:runLength,(length(waterLayerDepth)-bdepth):33) =
    0.71*p3;
    fwater(:,34) = 0.96;
    fwater(:,35) = 0.42;
end

```

```

% [HEAT CAPACITY AND THERMAL CONDUCTIVITY. TABLE 1]
latentHeat =334e6; % J/m3 water
Cice = 1.9e6; % heat capacity ice, J m-3 K-1
Corg = 2.5e6; % heat capacity organic material, J m-3 K-1
Cw = 4.18e6; % heat capacity liquid water, J m-3 K-1
Cair = 1.25e3; % heat capacity air, J m-3 K-1, wania = 1.25 e6
Cpeat = 3.44e4* fwater*100 + 583333; % heat capacity peat, J m-3 K-1,
Bonan 2002, with a regression to account for the percent saturation
Cmin = 2.00e6; % heat capacity mineral, J m-3 K-1, Wania 2009 Table 2,
Hillel 1982
Cvol= (fpeat.*Cpeat + fmin*Cmin + fair*Cair);
Cvol(:, 34)= 1.8e6;
Cvol(:, 35) = 2.7e6;

```

```

Lice = 2.2; % thermal conductivity ice, W m-1 K-1
Lorg = .25; % thermal conductivity organic matter, W m-1 K-1
Lw = 0.57; % thermal conductivity water, W m-1 K-1
Lair = 0.025; % thermal conductivity air, W m-1 K-1
Lmin = 2.00; % thermal conductivity mineral soil, W m-1 K-1, Wania, Table
2;
Lpeat = 0.06; % thermal conductivity peat soil, W m-1 K-1, 0% saturation;
Bonan 2002

```

```

Cond_Fr = cf*(fair*Lair + (1-fair).*(Lice.^((fwater)./(1-
fair))).*(Lpeat.^(fpeat./(1-fair))).*(Lmin.^(fmin./(1-fair))));
Cond_Th = ct*(fair*Lair + (1-fair).*(Lw.^((fwater)./(1-
fair))).*(Lpeat.^(fpeat./(1-fair))).*(Lmin.^(fmin./(1-fair))));
Cond_Th(:,34)= 2.12;
Cond_Th(:,35)= 2.16;
Cond_Fr(:,34)= 2.54;
Cond_Fr(:,35)= 2.51;
Cond_Fr(:,1)= .4;
Cond_Th(:,1)= .4;

```

```

heatCapacityOut=zeros(NumberOfSoilComputationNodes, runLength);
thermalConductivityOut=zeros(NumberOfSoilComputationNodes,
runLength);

```

```

% [SOLVE FOR SOIL TEMPERATURE NUMERICALLY]
for t = 1:runLength
soilTemp=zeros(1, NumberOfSoilComputationNodes);
    U1=zeros(1, NumberOfSoilComputationNodes);
    % // soil properties
        activeLayerDepth=0;
        thawingDepth=0;

% [initial conditions]
if t == 1.0
    prevSoilTemp = T_init;
    soilTemp=T_init;
    U1=prevSoilTemp;
end

% [burn conditions]
if t== burnDay
    soilTemp = horzcat(airTemp(burnDay),
    prevSoilTemp(bdepth:NumberOfSoilComputationNodes),
    prevSoilTemp((NumberOfSoilComputationNodes-bdepth +
3):NumberOfSoilComputationNodes));

else
    U1=prevSoilTemp;

S=TStep*60*60;%// ! 24 hours time step in seconds
    iter = 0;

```

```

while (iter < iter0) %&& maxABS > E0)
% // ! computation of boundary coefficients G1,G2
L0=soilThermalConductivity(soilLayerDepth(1),U1(1));
L1=soilThermalConductivity(soilLayerDepth(2),U1(2));

H0=soilLayerDepth(2)-soilLayerDepth(1);

if snowDepth (t) < E0 || prevSoilTemp(1) > E0
G1=0.0;
G2= airTemp(t);

elseif prevSoilTemp(1) <= 0.0 && snowDepth(t) < E0
G1=0.0;
G2=airTemp(t);

else%{ //! if (snowDepth(j) > 0.) then
ALFA = snowProperties(snowDensity, snowDepth(t));
ALFA= 1/ALFA;

C1=heatCapacityDynWT(soilLayerDepth(1),prevSoilTemp(1));
W1= 0.5* (L0+L1);
W2=H0*ALFA/W1;
W1=0.5*power(H0,2)*C1/W1/S;
G1=1.0 +W1+W2;
G2=(W2*airTemp (t) +W1*prevSoilTemp(1))/G1;
G1= 1/G1;
end

% // !---- Permutation and forward elimination
P1(2)=G1;
Q1(2)=G2;

for i=2:1:(NumberOfSoilComputationNodes-1)

C1=heatCapacityDynWT(soilLayerDepth(i),prevSoilTemp(i));

L2=soilThermalConductivity(soilLayerDepth(i+1),prevSoilTemp(i+1));
H1=soilLayerDepth(i+1)-soilLayerDepth(i);
H2=0.5*(H0+H1);
A1=0.5 * (L0+L1)*S/C1/(H0*H2);
B1= 0.5 *(L1+L2)*S/C1/(H1*H2);
C0= 1.0+A1+B1;

```

```

        P1(i+1)=B1/(C0-A1*P1(i));
        Q1(i+1)=(A1*Q1(i)+prevSoilTemp(i))*P1(i+1)/B1;
        H0=H1 ;
        L0=L1 ;
        L1=L2;

        heatCapacityOut(i,t)= C1;
        thermalConductivityOut(i,t) = L2;
    end

%
% //      ! computation of the Lower boundary koef. G3 & G4

C1=heatCapacityDynWT(soilLayerDepth(NumberOfSoilComputationNodes),
prevSoilTemp(NumberOfSoilComputationNodes));
G3= 0.5*power(H1,2)*C1/L2/S ;

G4=H1*G0+G3*prevSoilTemp(NumberOfSoilComputationNodes);
G3=1.0/(1.0+G3);
G4=G4*G3;

% //      ! Temperature computation in the last (deepest) grid
node

W1=(G3*Q1(NumberOfSoilComputationNodes)+G4)/(1.0-
G3*P1(NumberOfSoilComputationNodes));
maxABS=abs(W1-U1(NumberOfSoilComputationNodes));
U1(NumberOfSoilComputationNodes)=W1;

%
% //      !---- Back substitution
i= (NumberOfSoilComputationNodes-1);
while (i>=1)
%      { //DO WHILE (i>=1)
        W1=P1(i+1)*U1(i+1)+Q1(i+1);
% //      ! check for the iterative convergence
        if (abs(W1-U1(i))> maxABS)
            maxABS=abs(W1-U1(i));
        end
        U1(i)=W1;

        i=i-1;
    end      %[//ENDDO !WHILE]

```

```

    iter=iter+1;
    end          %[end while ((ITER < ITER0).AND.(maxABS > E0))]

soilTemp = U1 ;
end

%[ITERATE SOIL TEMPERATURES, CREATE OUTPUT]
prevSoilTemp=soilTemp;
soilTempOut(1:NumberOfSoilComputationNodes, t) = soilTemp.;
activeLayerOut(1, t)= activeLayerDepth;
thawLayerOut(1,t) = thawingDepth;
soilT1 =soilTemp(1);
unfrozWaterContent(1:NumberOfSoilComputationNodes, t)=
unfrozenWaterContent(soilTempOut(t), soilLayerDepth);

end

% [WRITE OUTPUT FILES]
soilTempOut1= horzcat(soilLayerDepth.', soilTempOut);
csvwrite([filer '.dat'], soilTempOut1)
csvwrite([filer '_HC.dat'], heatCapacityOut)
csvwrite([filer '_TC.dat'], thermalConductivityOut)
csvwrite([filer '_fwaterVol.dat'], fwater)

```

FUNCTIONS

VOLUMERIC HEAT CAPACITY

```
function CAP= heatCapacityDynWT (depth, temper)
Tfr = evalin('base', 'Tfr'); % Temperature of freezing
FIT = evalin('base', 'FIT'); % "corridor of phase change, 0 degrees +/- FIT. FIT
= 0.08 degrees
Cvol = evalin('base', 'Cvol'); % volumetric heat capacity of things other
than water
fwater = evalin('base', 'fwater'); % total amount of water (ice + liquid
water)
latentHeat = 334e6; % (J m-3 k-1)
Cw = evalin('base', 'Cw');
Cice = evalin('base', 'Cice');
tempRange= 0.5;%2*FIT;
Wunf= evalin('base', 'Wunf');

t = evalin('base', 't');
j = soilNumLayers(depth);

    if (temper < Tfr - 0.5)%tempRange)
        CAP=Cvol(t, j)+ Cice*(fwater(t, j)-Wunf) +Cw*Wunf;
    elseif (temper > Tfr)%+ tempRange)
        CAP=Cvol(t, j)+Cw*fwater(t, j);
    else
        CAP= Cvol(t,j) + 0.5*Cice* ( fwater(t,j)- Wunf )+ 0.5*Cw*
        (fwater(t,j)- Wunf ) + latentHeat*fwater(t,j)/tempRange;
    end
end
```

SOIL THERMAL CONDUCTIVITY

```
function COND = soilThermalConductivity(depth,temper)
Tfr = evalin('base', 'Tfr'); % Temperature of freezing
FIT = evalin('base', 'FIT'); % "corridor of phase change, 0 degrees +/- FIT. FIT
= 0.08 degrees
Cond_Fr = evalin('base', 'Cond_Fr'); % thermal conductivity frozen soil
Cond_Th = evalin('base', 'Cond_Th'); % thermal conductivity thawed soil
t= evalin('base', 't');
    i=soilNumLayers(depth);
    if i==0
        i=1;
    end

    if (temper < - 0.5 )%Tfr -FIT)
        COND=Cond_Fr(t, i);
    elseif (temper > 0)%Tfr -FIT)
        COND= Cond_Th(t, i);
    else
        COND= 0.5 *(Cond_Th(t, i)+Cond_Fr(t, i));
    end
end
```

UNFROZEN WATER CONTENT

```
function unfrWater= unfrozenWaterContent(temper, depth)
a = evalin('base', 'a');
b = evalin('base', 'b');
c = evalin('base', 'c');

    i=numLayers(depth);
    ac = a(i);
    bc=b(i);
    cc=c(i);
    cc = 0.0;
    unfrWater=ac * power(abs(cc-temper),bc);
end
```

```

NUMLAYERS
function Numl= numLayers(depth)
    i=1;
    soilLayerThickness= evalin('base', 'soilLayerThickness');
    if depth > soilLayerThickness(i) + 1e-3
        i=i+1;
    end
    Numl=i;
end

```

```

SOIL NUMLAYERS
function Numl = soilNumLayers(depth)
% determines which soil layer each water layer falls into, and therefore
% which soil properties each layer has

numSoilLayers = evalin('base', 'numSoilLayers');
soilLayerThickness = evalin('base', 'soilLayerThickness');

    for i0 = 2:1:(numSoilLayers) % find the bottom depth of each layer.
        soilLayerThickness(i0)=soilLayerThickness(i0)+
            soilLayerThickness(i0-1);
    end

    i= 1;
    for j = 1:1:length(soilLayerThickness)
        if depth > soilLayerThickness(j) + 1e-3
            i=i+1;
        end
        if i > numSoilLayers
            i = numSoilLayers;
        end
    end
    Numl=i;
end

```


WATER CONTENT IN PEAT LAYERS

1. PEAT LAYER 1 (FIBRIC/SURFACE)

```
function WC= nearSurfWat(depth, WT)
% this function determines the amount of water at a surface peat site
% given a depth and water table depth. Relationships described from
% TDR at Mer Bleue that Steve sent.

    distance = WT - depth;
    if distance < 0.1
        WC = 1 - distance*9;
    else
        WC = 0.1 - distance * 0.1;
    end

    if WC <0.03
        WC = 0.03;
    end
end
```

2. PEAT LAYER 2 (MESIC)

```
function WC= mesWat(depth, WT)
% this function determines the amount of water at a surface peat site
% given a depth and water table depth. Relationships described from
% TDR at Mer Bleue that Steve sent.

    distance = WT - depth;
    if distance < 0.1
        WC = 1 - distance*5;
    else
        WC = 0.5 - distance * 0.6;
    end

    if WC <0.03
        WC = 0.03;
    end
end
```

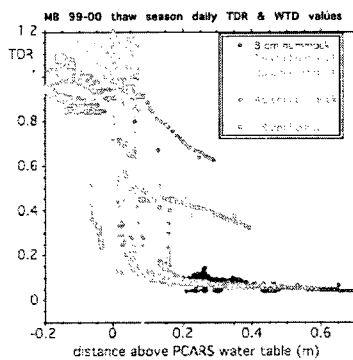
3. PEAT LAYER 3 (HUMIC)

function WC= deepWat(depth, WT)

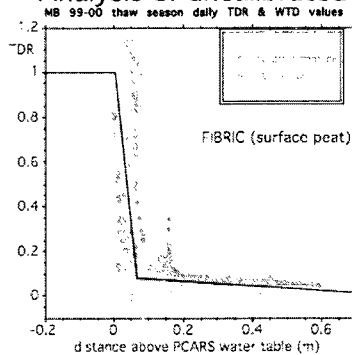
% this function determines the amount of water at a surface peat site
 % given a depth and water table depth. Relationships described from
 % TDR at Mer Bleue that Steve sent.

```

distance = WT - depth;
WC = 1 - distance * 1.33;
if WC < 0.03
    WC = 0.03;
end
end
end
  
```

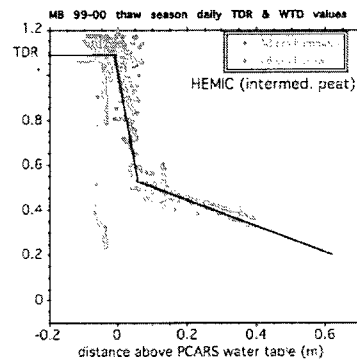


Analysis of uncalibrated TDR data from Mer Bleue

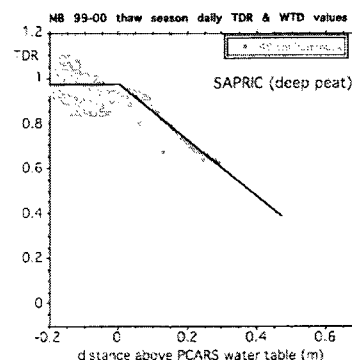


Peat water content vs.
distance above WT.

near surface -
rapid dewatering



mid-level -
moderate dewatering



deep peat -
slow dewatering