

**Exploring the
sensitivity of soil C
dynamics**

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Exploring the sensitivity of soil carbon dynamics to climate change, fire disturbance and permafrost thaw in a black spruce ecosystem

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Abstract

In the boreal region, soil organic carbon (OC) dynamics are strongly governed by the interaction between wildfire and permafrost. Using a combination of field measurements, numerical modeling of soil thermal dynamics, and process-based modeling of OC dynamics, we tested the sensitivity of soil OC storage to a suite of individual climate factors (air temperature, soil moisture, and snow depth) and fire severity. We also conducted sensitivity analyses to explore the combined effects of fire-moisture interactions and snow seasonality on OC storage. OC losses were calculated as the difference in OC stocks after three fire cycles (~ 450 years) following a prescribed step-change in climate and/or fire. Across single-factor scenarios, our findings indicate that warmer air temperatures resulted in the largest soil OC losses (5.3 kg C m^{-2}), whereas dry soil conditions alone (in the absence of wildfire) resulted in the smallest carbon losses (0.1 kg C m^{-2}). Increased fire severity resulted in carbon loss of 3.3 kg C m^{-2} , whereas changes in snow depth resulted in smaller OC losses ($2.1\text{--}2.2 \text{ kg C m}^{-2}$). Across multiple climate factors, we observed larger OC losses than for single-factor scenarios. For instance, high fire severity regime associated with warmer and drier conditions resulted in OC losses of 6.1 kg C m^{-2} , whereas a low fire severity regime associated with warmer and wetter conditions resulted in OC losses of 5.6 kg C m^{-2} . A longer snow-free season associated with future warming resulted in OC losses of 5.4 kg C m^{-2} . Soil climate was the dominant control on soil OC loss, governing the sensitivity of microbial decomposers to fluctuations in temperature and soil moisture; this control, in turn, is governed by interannual changes in active layer depth. Transitional responses of the active layer depth to fire regimes also contributed to OC losses, primarily by determining the proportion of OC into frozen and unfrozen soil layers. Carbon cycle feedbacks from the boreal region to the climate system will clearly depend upon these interactions among climate drivers, fire regime characteristics, and permafrost dynamics.

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1 Introduction

High-latitude soils store large quantities of organic carbon (OC), accounting for nearly 50% of the global belowground OC pool (Tarnocai et al., 2009). Accumulation of soil OC in many northern soils is governed by the presence of permafrost, which functions as a large reservoir for OC (Zimov et al., 2006). Recent warming at high latitudes has caused localized thawing of permafrost (Osterkamp and Romanovsky, 1999; Romanovsky et al., 2010), resulting in the release of old OC from terrestrial ecosystems (Schuur et al., 2009). Permafrost thaw also can alter local soil thermal and hydrologic conditions (Jorgenson et al., 2001; Yi et al., 2009b), which can influence rates of microbial activity and OC loss from decomposition. Modeling analyses project widespread thawing of permafrost in the next century (Euskirchen et al., 2006; Lawrence et al., 2008), but considerable uncertainty exists regarding the fate of soil OC following thaw (Limpens et al., 2008; Schuur et al., 2008).

Wildfire has the potential to exacerbate rates of permafrost thaw (Yoshikawa et al., 2003) and OC losses (Harden et al., 2000) from soils of the boreal region. Through the combustion of surface organic horizons, wildfire reduces thermal insulation and increases active layer depth (ALD; Burn, 1998; Yoshikawa et al., 2003; Johnstone et al., 2010). Re-accumulation of the organic horizon between fire cycles can allow for the recovery of ALD to pre-fire depths (Yi et al., 2010). This interaction between wildfire and permafrost can have a profound influence on soil OC accumulation and loss (O'Donnell et al., 2010). ALD fluctuations across fire cycles determine the proportion of OC in unfrozen or frozen mineral soil, and thus, rates of OC accumulation or loss in deep soil horizons (O'Donnell et al., 2010). In this manner, ALD functions as a control on substrate (i.e., OC) availability to microbial decomposers in unfrozen soils. Furthermore, the magnitude of the ALD effect on soil OC loss is closely linked to fire regime characteristics, such as fire severity (O'Donnell et al., 2010).

Soil OC dynamics are also governed by the sensitivity of microbial decomposition to changes in soil climate, which in the boreal region, is governed by complex interactions

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among physical drivers, soil properties and disturbance (Jorgenson et al., 2010). For example, increased soil temperatures following fire can stimulate decomposition rates (Richter et al., 2000; O'Neill et al., 2002). However, the temperature sensitivity of decomposition may be reduced due to post-fire changes in soil moisture or OC quality (Neff et al., 2005; O'Donnell et al., 2009b). Wildfire can increase soil moisture content via reduced rates of evapotranspiration and interception (Moody and Martin, 2001), but can also decrease following fire (O'Neill et al., 2002). Similarly, permafrost thaw has caused both wetting and drying of terrestrial ecosystems, depending upon topographic position and ground ice content (Jorgenson et al., 2001; Jorgenson and Osterkamp, 2005). Increased snow depth can enhance soil warming and permafrost thaw (Nowinski et al., 2010), whereas decreases in snow-depth can promote rapid freezing during cooling periods (Goodrich, 1982; Stieglitz et al., 2003). The relative sensitivity of soil thermal conditions to these climatic drivers and disturbances will ultimately govern rates of decomposition, and in turn, will determine OC storage across the boreal landscape.

Here, we build on prior field and modeling studies (Harden et al., 2000; O'Donnell et al., 2010) to assess the relative sensitivity of ALD and soil climate effects on soil OC storage in a black spruce ecosystem of Interior Alaska. To achieve this objective, we used a combination of field measurements of soil temperature, moisture and snow dynamics, numerical modeling of soil thermal dynamics, and process-based modeling of soil OC dynamics. Using this approach, we addressed the following questions. (1) What is the sensitivity of ALD and soil climate to air temperature, soil moisture, snow depth, and fire severity? (2) What is the sensitivity of soil OC storage to air temperature, soil moisture, snow depth, and fire severity? (3) What are the relative effects of ALD, soil climate, and ALD-soil climate interactions on soil OC losses across climate and fire scenarios? By addressing these questions, we aim to refine our understanding of the mechanisms underlying soil OC dynamics in relation to wildfire and permafrost thaw in the boreal region.

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2 Methods

2.1 Experimental design

Previous work demonstrated the fundamental importance of active layer depth (ALD) (O'Donnell et al., 2010) and soil temperature (Carrasco et al., 2006) in governing the net C balance of boreal soils. For this study, we combined field observations of temperature and soil moisture, numerical modeling, and process-based modeling to quantify the relative effects of individual and combined climate/fire factors on net soil OC accumulation and loss (Fig. 1; Table 1). In this study, we modified the Fire-C model (Harden et al., 2000; O'Donnell et al., 2010) to improve simulations of ALD and OC dynamics in order to (1) account for deeper permafrost OC stocks, and (2) to incorporate in more detail the temperature and moisture controls on decomposition. In previous work, thermal conductivity of the organic horizon was measured across a range of moisture contents (O'Donnell et al., 2009a) and ALD was calculated as a function of organic horizon thickness across an upland fire chronosequence near Hess Creek, AK (see O'Donnell et al., 2010). For this study, we used the Geophysical Institute Permafrost Laboratory (GIPL) model (Romanovsky and Osterkamp, 1997; Nicolsky et al., 2007; Marchenko et al., 2008) to assess the sensitivity of ALD and soil temperature to a wide range of climatic factors (air temperature, moisture, snow, and fire severity) both as separate and combined. We then used output from these sensitivity analyses to assess the effects of ALD and soil climate on organic matter decomposition and soil OC accumulation using a simple process-based, long-term model (O'Donnell et al., 2010).

2.2 Modeling soil carbon dynamics

The soil OC dynamics model used in this study was originally developed by Harden et al. (2000) to track long-term trends in soil carbon inventories of shallow and deep organic horizons across fire cycles. This model was subsequently modified to include inputs from coarse woody debris (Manies et al., 2005), to operate on annual or monthly

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time-steps (Carrasco et al., 2006), and to include OC dynamics in mineral soil of the active layer and near-surface permafrost (O'Donnell et al., 2010). Briefly, the model examines net changes in soil OC storage over time (dC/dt) as governed by OC inputs (net primary production (NPP) from moss and trees) and losses (heterotrophic respiration and fire). Here we discuss further improvements and modifications to the model, which include GIPL-modeled active layer dynamics, and temperature and moisture regulation of decomposition.

2.2.1 Active layer dynamics

In the modeling study by O'Donnell et al. (2010), active layer depth was calculated simply as a function of organic horizon thickness at time t (in years) for the study area near Hess Creek. To more precisely represent active layer dynamics in the model, we reconstructed active layer depth vs. organic horizon thickness regression curves across a range of climate conditions using the GIPL model (as discussed in Sect. 2.4). The equations derived from these relationships were then used to inform the soil OC dynamics model across different climatic scenarios.

2.2.2 Deep soil carbon

In a prior version of the model, we tracked soil C inventories between the ground surface and a depth of one meter (O'Donnell et al., 2010). Here, we have parameterized the model to track soil OC inventories in the top 2 m. In most mature forests in Alaska, soil OC between one and two meters resides in the permafrost pool (Tarnocai et al., 2009). However, post-fire thawing of near-surface permafrost can result in thaw depths exceeding one meter (e.g., Viereck et al., 2008), which might impact deeper OC stocks (>1 m).

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2.2.3 Temperature and moisture controls on decomposition

In a prior version of the model (O'Donnell et al., 2010), decomposition constants (k) were derived using chronosequence and radiocarbon methodologies. In that study, k reflected the apparent decomposability of soil organic matter (see Davidson and Janssens, 2006), integrating the effects of temperature, soil moisture and organic matter quality on decomposition during historic times. However, future climate scenarios require that k respond to fluctuations in soil temperature and moisture.

We used the following equation to calculate an inherent k , which integrates the decomposability of a substrate based on its molecular structure at ambient temperature:

$$K_{\text{base}} = ke^{rT} \times f(\theta) \quad (1)$$

where K_{base} is the apparent decomposition constant as reported in O'Donnell et al. (2010), k is the inherent decomposition constant, r is a parameter describing the temperature sensitivity of decomposition, T is the mean monthly soil temperature for the month of August (as measured or calculated across the fire chronosequence), when maximum thaw depth occurs during the year, and $f(\theta)$ is a soil moisture scalar that depends on gravimetric moisture content. The function $f(\theta)$ is a value between 0 and 1, with optimal decomposition occurring when $f(\theta)$ is around 0.6, and was calculated following the equations of Frohling et al. (1996) for a feather moss-dominated black spruce forest. To determine a value for r , we used a temperature sensitivity quotient (Q_{10}) equation:

$$Q_{10} = e^{r10} \quad (2)$$

We used a different Q_{10} value for each soil layer, as reported in the literature. For shallow organic horizons, we prescribed a Q_{10} of 1.9, as reported by Wickland and Neff (2007) for fibric organic matter. For deep organic horizons (i.e., mesic or humic horizons), we prescribed a Q_{10} value of 4.4, as reported by Dioumaeva et al. (2003). For unfrozen mineral soil of the active layer, we prescribed a Q_{10} of 2.0, as reported in other studies (Clein and Schimel, 1995; Frohling et al., 1996; Carrasco et al., 2006).

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For near-surface permafrost, we lowered the Q_{10} value of unfrozen mineral soil to 30% of its value, following the work of Dioumaeva et al. (2003) and Fan et al. (2008), to account for reduced microbial activity at temperatures below freezing. We then used soil temperature output from the GIPL model (see *Modeling soil thermal dynamics* section for details) to calculate T across all climate and fire sensitivity analyses (Fig. 1). $f(\theta)$ was only re-calculated for the two climate scenarios in which we prescribed wetter or drier conditions.

2.3 Field measurements of soil temperature, moisture and snow depth

We measured soil temperature, soil moisture, and ALD and snow depth at three stand ages (unburned mature, 2003 burn, 1967 burn) from the fire chronosequence of O'Donnell et al. (2010). These different stand ages provide a natural gradient in soil properties such as organic horizon thickness, soil thermal dynamics, ALD, and soil OC stocks. From September 2007 to September 2009, we monitored soil temperature every 2 h using HOBO Pro V2 two-channel dataloggers (Onset Computer Corporation, Pocasset, MA, USA). We also used HOBO Smart Temperature sensors to monitor temperature at a single depth at each site. Temperature probes were installed at the ground surface, at the top of the permafrost (i.e., TTOP), near 2 m below the ground surface, and also at the interface of key organic soil horizons. Soil moisture of organic soil horizons was also logged every 2 h at each stand age using ECH₂O Smart Soil Moisture probes routed to a HOBO microstation (Onset Computer Corporation, Pocasset, MA, USA). Soil moisture probes were installed within distinct organic soil horizons. In addition to temperature and soil moisture, we conducted snow surveys over two winters. At each stand age, we measured snow depth every 6 m across a 30 m linear transect on 13 November, 25 February, and 3 April 2007 and 2008 and we also measured snow depth on 2 April 2009.

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2.4 Modeling soil temperature dynamics

We used the GIPL model to simulate the depth of seasonal freezing and thawing by solving a one-dimensional non-linear heat equation with phase change (Nicolosky et al., 2007). To calibrate the GIPL thermal model for the Hess Creek study sites, we combined field measurements of soil temperature (Supplement Figs. 1 and 2), soil moisture (Supplement Fig. 3), ALD, snow depth, field descriptions of soil horizon thickness (O'Donnell et al., 2010), and empirical calculations of thermal conductivity (O'Donnell et al., 2009a). For each study site (unburned mature, 2003 burn, 1967 burn), the measured soil temperature data were used to calibrate the model, which in turn was used to reconstruct the one-dimensional temperature field in the active layer and near-surface permafrost. We used air temperature and snow depth measurements from Hess Creek to drive model simulations of soil temperature. For soil parameterizations, we divided the organic horizon into three layers: live/dead moss, fibrous organic matter, and amorphous organic matter (Table 2), following the approach of Yi et al. (2009a). Mineral soil horizons were divided into active layer soil, frozen loess, and frozen bedrock. We prescribed soil moisture content for organic horizons based on mean volumetric water content (VWC) measured during the summer of 2008 (Table 3; Supplement Fig. 3). Unfrozen thermal conductivity values for organic horizons were calculated from measured VWC values (O'Donnell et al., 2009a; Table 3). Frozen thermal conductivity, volumetric heat capacity, and volumetric unfrozen water content parameters were prescribed following previous studies (Romanovsky and Osterkamp, 1997, 2000).

Following model calibration, we used the GIPL model to determine the sensitivity of ALD and soil temperatures to changes in air temperature, snow dynamics, soil moisture, and fire severity (Fig. 1; Table 1). For all sensitivity analyses, we introduced a step-change in climate or fire disturbance in order to evaluate effects on soil thermal conditions at the year 2100 relative to the present day. Using the step-change approach, we compared the importance of various factors in governing soil thermal dynamics in a black spruce ecosystem. Climate and fire factors were considered both

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in isolation and as combined effects. These simulations were designed to explore the underlying mechanisms controlling OC loss from boreal soils, and not to provide an exact estimate of future permafrost thaw. For a more precise estimation of ALD and soil temperature responses to future climate and fire, a transient simulation from the present day to 2100 (e.g., Lawrence et al., 2008) would be more appropriate.

To assess the effect of atmospheric warming, we increased mean daily air temperature (as measured at Hess Creek) to match predicted air temperatures under the A2 (business as usual) and B1 (reduced greenhouse gas emissions) warming scenarios (Christensen et al., 2007; Fig. 2). To assess the effects of future wetting or drying scenarios on active layer depth, we increased and decreased soil moisture content of organic horizons by 25% relative to present-day measurements (Supplement Fig. 3), with concomitant changes in thermal conductivity following O'Donnell et al. (2009a). We also ran two alternate snow scenarios where daily snow depth was increased by 25% and decreased by 25% relative to present-day measurements. To assess the sensitivity of ALD and soil temperatures to fire severity, we first prescribed a moderate fire severity (64% of organic matter combusted) and high fire severity (77% of organic matter combusted, following Kane et al., 2007) relative to present-day fire severity (41% of organic matter combusted; Harden et al., 2006).

We also ran scenarios to test the combined effects of climate and fire severity. For instance, we ran two scenarios to test the interactive effects of fire severity (high, low), moisture (wet, dry) and warming (A2 scenario; Table 1) relative to the present day. We also ran a scenario to reflect changes in snow seasonality associated with future warming. Following estimates of temporal snow dynamics by Euskirchen et al. (2009), our simulations reflected snow conditions at the year 2100, with snowmelt occurring 21 days earlier in the spring and snow return occurring 20 days later in the fall as compared present-day conditions. The A2 warming scenario was also used as driving data for the changing snow seasonality treatment.

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2.5 Sensitivity of soil carbon to future climate and fire scenarios

To assess the sensitivity of soil OC to future climate and fire scenarios, we used output (ALD, soil temperature) from the GIPL model to drive OC storage and decomposition in the Fire-C model (Fig. 1). Other studies have coupled thermal and ecosystem models to investigate soil OC dynamics in northern ecosystems (Zhuang et al., 2001, 2002; Khvorostyanov et al., 2008; Yi et al., 2009b). Our approach here is unique in that it incorporates field and laboratory parameterizations (O'Donnell et al., 2009a, 2010) into a modeling framework in which we explicitly evaluate fire-permafrost interactive effects on soil OC across a range of future climate conditions. At this stage, the model does not consider the effects of talik development on decomposition of soil organic matter. For the model spin-up (0–6500 years), we ran the Fire-C model following parameterizations of O'Donnell et al. (2010), which reflect historic climatic conditions and fire regimes. At the year 6500, we modified Fire-C parameters to reflect future climatic conditions and fire regimes. We tested the sensitivity of soil OC to air temperature (A2 vs. B1 scenarios), soil moisture (wetter vs. drier), snow depth (deeper vs. shallower), and fire severity (low severity, or 25% vs. high severity, or 77%). We also tested the sensitivity of soil OC to the combined climate and fire effects. Specifically, we evaluated the effects of increased fire severity associated with drier and warmer condition, decreased fire severity with wetter and warmer conditions, and changes in snow seasonality associated with future warming. The net change in OC storage for each scenario was calculated as the difference between total OC stocks (top 2 m) at 6500 (i.e., present day) and 7000 years (i.e., after 3 fire cycles).

Model simulations were run to assess the magnitude of both ALD and soil climate effects on soil OC dynamics. To isolate the ALD effect on soil OC, we first ran the Fire-C model by only changing the relationship between ALD and organic horizon thickness, as simulated by the GIPL model. To isolate the soil climate effect on soil OC, we then ran the Fire-C model by only changing the soil temperature (as simulated by GIPL) and moisture controls on decomposition rates. To assess the interactive effects of ALD

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and soil climate on soil OC, we ran the Fire-C model a third time changing both ALD dynamics and soil climate controls on decomposition rates. For each effect (ALD, soil climate, ALD+soil climate), we calculated the net effect on soil OC accumulation or loss.

3 Results

3.1 Thermal model calibration

Field measurements of air temperature and snow depth were used to calibrate the GIPL model. Air temperature varied seasonally (black line in Fig. 2), with maxima occurring during July (monthly mean=14.9°C) and minima during January (monthly mean=-25.3°C). Snow depth varied across survey dates but not among study sites across the fire chronosequence (Fig. 3a). Minimum snow depths were measured in November, ranging between 7 and 9 cm across study sites. Maximum snow depths were measured in April and varied between years. In winter 2007–2008, maximum snow depths averaged between 41 and 48 cm, whereas in winter 2008–2009, maximum snow depths averaged between 60 and 67 cm. We observed a marginally good relationship between snow depths at Hess Creek and at a long-term monitoring site near Fairbanks, AK (Fig. 3b), described by the equation: $y = -4.0822 + 0.8502x$ ($R^2 = 0.78$; $P = 0.12$). We used this relationship to calculate daily snow depth at Hess Creek during model calibration.

We observed good agreement between measured and modeled soil temperatures (Fig. 4) and ALD (Supplement Fig. 4) at sites across the Hess Creek fire chronosequence. At the ground surface, modeled mean daily temperature (MDT) closely tracked observed MDT during winter, summer and during seasonal transitions (e.g., snowmelt). Deeper in the soil profile, we observed small differences (<0.5°C) during winter and summer between modeled and observed ground temperatures (Fig. 4b–d). The differences were most pronounced during periods of cooling in the fall/early winter and

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warming during spring/early summer. The differences in ground temperature during periods of cooling have been attributed to the influence of unfrozen water content (Romanovsky and Osterkamp, 1997, 2000). During periods of warming in spring, ground temperatures were likely influenced by infiltration and re-freezing of snowmelt waters.

5 These processes typically result in the formation of a “zero-degree curtain” at the ground surface, which we observed at our study sites during spring. Following the completion of snowmelt, thawing of the active layer began with ground temperatures $>0^{\circ}\text{C}$. Modeled ALD was within 3 cm of the observed ALD across the three study sites (Supplement Fig. 4).

10 **3.2 Modeling the sensitivity of active layer depth and soil climate to climate and fire**

Warmer air temperatures alone had a large impact on ALD (Fig. 5a; Table 4), with the A2 scenario increasing ALD by an average of 37 ± 3 cm relative to present-day air temperatures across a range of organic horizon thicknesses. The B1 scenario resulted in considerably smaller increases in ALD, averaging 9 ± 1 cm across a range of organic horizon thicknesses. Changes in soil moisture had a moderate influence on ALD (Fig. 5b), with wetter conditions driving an average ALD increase of 5.4 ± 0.7 cm and drier conditions driving an average ALD decrease of 7.0 ± 1.2 cm across a range of organic horizon thicknesses. Snow depth had a very small impact on ALD, with average ALD increasing or decreasing by 2 ± 0 cm with deeper or shallow snowpack, respectively (Fig. 5c). High fire severity caused an average increase in ALD of 18 ± 5 cm, whereas low fire severity caused an average decrease in ALD of 7 ± 1 cm (Fig. 5d).

In addition to driving changes in ALD, individual climate and fire factors caused changes in soil temperature in the active layer and near-surface permafrost (Fig. 6). Warming associated with the A2 scenario increased soil temperature by $3.5\text{--}4.5^{\circ}\text{C}$ for August MMT in active layer soils. The B1 scenario resulted in higher soil temperatures by 0.7 to 1.0°C relative to present-day climate conditions. Soil moisture also had a moderate influence on August MMT in the active layer, as wetter conditions resulted

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in an increase of 0.3–0.6 °C and drier conditions resulted in decrease of up to 0.5 °C relative to present-day conditions. Deeper snowpack resulted in slight increases and shallower snowpack resulted in slight decreases in August MMT relative to present-day conditions. Increased fire severity (as prescribed through changes in organic horizon thickness) increased August MMT by 0.4 to 1.5 °C, whereas decreased fire severity cooled August MMT by up to 0.3 °C relative to present-day conditions.

When considered as combined effects, we found that high fire severity, drier soils and warmer air temperatures resulted in the largest increase in ALD (mean = 56 ± 6 cm; Fig. 7a; Table 4) and August MMT (4.7–5.5 °C increase in active layer soils; Fig. 7b). The combined effects of low severity fires, wetter soils and warmer air temperatures resulted in more modest increases in ALD (mean = 37 ± 4 cm) and August (3.6–5.0 °C increase in active layer soils). Longer snow-free seasons associated with future warming also resulted in large increases in ALD (mean = 39 ± 3 cm) and August MMT (3.6–4.6 °C increase in active layer soils).

3.3 Modeling the sensitivity of soil carbon to climate and fire

Of the individual climate factors, the A2 warming scenario (no change in fire severity) resulted in the largest OC losses from soil (5.3 kg C m⁻²; Fig. 8), whereas the drier soil moisture scenario in the smallest losses (0.06 kg C m⁻²). After the A2 scenario, the high fire severity treatment resulted in the next largest OC losses (3.3 kg C m⁻²). Changes in snow depth and increases in soil moisture resulted in minimal OC losses, ranging from 2.1–2.2 kg C m⁻². When the model was run for three fire cycles under the present-day climate (with no changes in snow, moisture or fire), we also observed an OC loss of 2.1 kg C m⁻². Across individual climate scenarios, interannual changes in ALD (i.e., the ALD effect) accounted for 61% of the total OC losses, whereas changes in soil temperature and/or moisture (i.e., the soil climate effect) accounted for 82% of total OC losses.

When multiple climate factors were considered as combined effects, we generally observed larger OC losses from soil relative to the individual factor analyses (Fig. 9).

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The high fire severity scenario associated with drier soil and warmer air temperature resulted in a soil OC loss of 6.1 kg C m^{-2} . The low fire severity scenario associated with wetter soil and warmer air temperature resulted in a soil OC loss of 5.6 kg C m^{-2} . The scenario reflecting longer snow-free season and warmer air temperatures resulted in a soil OC loss of 5.4 kg C m^{-2} . Across combined-effect scenarios, the ALD effect accounted for 39% of total OC losses, while the soil climate effect accounted for 79% of total OC losses.

4 Discussion

4.1 Effects of air temperature on active layer depth, soil climate and soil carbon

Terrestrial ecosystems have the potential to act as a positive feedback to the climate system in response to warmer air temperatures (e.g., Friedlingstein et al., 2006), given the temperature sensitivity of heterotrophic respiration and subsequent CO_2 release from soils. In this study, we illustrate the importance of higher air temperatures for OC loss from boreal soil. When compared to other individual factors (soil moisture, snow depth, fire severity), higher air temperatures caused the largest OC losses from soil in an upland black spruce ecosystem. Through our modeling analyses, we were able to identify two mechanisms underlying OC losses. First, higher air temperatures resulted in warmer soil temperatures, which stimulated microbial decomposition of soil organic matter following Q_{10} value for each soil horizon. Second, higher air temperatures, and in particular the A2 warming scenario, resulted in large increases in active layer depth (ALD), which transferred considerable amounts of OC from the permafrost to a thawed C pool (e.g., Schuur et al., 2008), where decomposition rates were higher (Mikan et al., 2002). Above 0°C , decomposition rates are typically described by an exponential relationship with temperature (Lloyd and Taylor, 1994; Kirschbaum, 2000), whereas decomposition rates below 0°C are indirectly related to temperature, constrained by the availability of unfrozen water (Rivkina et al., 2000; Romanovsky and Osterkamp, 2000). This complex interaction between physical and biological processes has contributed

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to the ongoing uncertainty regarding the temperature sensitivity of decomposition and carbon cycle feedbacks to the climate system (Davidson and Janssens, 2006; Kirschbaum, 2006; Heimann and Reichstein, 2008). Explicit consideration of freeze-thaw fronts (e.g., Yi et al., 2009) and phase change effects on soil OC dynamics in ecosystem models will help in reducing these uncertainties. Furthermore, increased heterotrophic respiration associated with warming will likely be accompanied by increased nitrogen availability and plant production (Mack et al., 2004), although the net C balance is likely to favor an increase of CO₂ relative to pre-disturbance conditions (Bond-Lamberty et al., 2007).

4.2 Effects of fire severity versus warming

In a prior study, O'Donnell et al. (2010) illustrated the importance of fire severity on soil OC loss from deep soil horizons using the Fire-C model. Here, we show that the direct effects of increased fire severity on soil OC storage, while substantial, are generally smaller than the direct effects of warmer air temperatures. In reality, greater fire severity always results in warmer soils because of the insulating effect of organic layers on soil temperatures and active layer depth. For example, using the GIPL model, we showed that high severity fires result in increased ALD and soil temperatures following deep combustion of organic soil horizons (Fig. 5d), which enhanced OC losses from soil. These findings are consistent with other sensitivity analyses conducted in other black spruce ecosystems in the boreal region. In a recent analysis, Yi et al. (2010) illustrate that this post-fire soil warming and increases in ALD enhance soil decomposition and N mineralization in early successional stands. Zhuang et al. (2002) illustrated that soil OC stocks in black spruce ecosystems are highly sensitive to fire severity, with large OC losses associated with high severity burns. Therefore, as future warming will likely increase the size and severity of wildfires in the boreal region (Balshi et al., 2009; Flannigan et al., 2009), increases in OC loss from boreal soils will result.

We also showed that soil OC stocks were also sensitive to changes in soil moisture. Under drier soil conditions (in the absence of fire), we observed relatively low

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OC losses from soil, a response driven by both physical and biological factors. Decreased soil moisture reduced the thermal conductivity of organic soil horizons, which is a dominant control on ALD in black spruce ecosystems (Bonan, 1989; Yi et al., 2009b; Jorgenson et al., 2010). By reducing ALD, a larger proportion of OC stocks in mineral soil were reincorporated into the near-surface permafrost pool, where decomposition is essentially arrested. Together, cooler temperatures and drier conditions reduced decomposition rates, resulting in smaller OC losses relative to other scenarios. Under higher moisture conditions, we generally observed the opposite effect, with increased ALD, soil temperatures and decomposition.

Interactions between fire and soil drainage class will likely govern the extent of permafrost thaw and magnitude of soil OC losses following fire in the boreal region (Harden et al., 2001; Kane et al., 2007; Yi et al., 2009a). We conducted additional simulations to assess the combined effects of fire severity (high, low), soil moisture (wet, dry) and warming on permafrost and soil OC dynamics. Surprisingly, we observed a similar magnitude of OC loss from both the high severity-dry treatment and the low-severity-wet treatment. This finding is likely due to a shift in the dominant mechanism driving soil temperature and OC dynamics between treatments. For instance, under the low severity-wet treatment, soil OC dynamics appear to be more sensitive to increased soil moisture and heat transfer than to the large reduction in organic horizon thickness than the decrease in soil moisture. In contrast, under the high severity-dry treatment, soil OC dynamics appear to be more sensitive to the large reduction in organic horizon thickness than to the decrease in soil moisture. Prior studies have documented the importance of organic horizon thickness and moisture on soil temperatures at the ground surface (Harden et al., 2006) and in near-surface permafrost (Jorgenson et al., 2010). The response of OC and permafrost dynamics in black spruce ecosystems to future climate and fire regimes will likely depend on these feedbacks among organic horizon thickness, soil drainage, and soil temperature and permafrost dynamics (Johnstone et al., 2010).

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The interactive effects of fire and soil moisture on soil OC dynamics may depend on localized conditions relating to parent material and ground ice content (Jorgenson et al., 2010). Across the fire chronosequence at Hess Creek, soil moisture increased in surface soil horizons immediately following fire and persisted for at least 40 years after the burn (Supplement Fig. 4). This observation is counter to the findings of Yoshikawa et al. (2003), who for a range of burned stands in Interior Alaska, observed higher moisture content immediately following fire but lower moisture content in maturing stands. Sites examined by Yoshikawa et al. (2003) were typically underlain by mica schist with a thin loess mantel (Rieger et al., 1972; Ping et al., 2005), whereas our study sites were underlain by thick loess deposits (O'Donnell et al., 2010). Furthermore, we observed cryostructures consistent with syngenetic permafrost and the common occurrence of an ice-rich intermediate layer (Shur et al., 2010), which mediates thaw of deep permafrost due to the high-latent heat content of ice (Shur et al., 2005). This ice-rich layer may have influenced post-fire soil conditions by both contributing thaw water to the active layer and by minimizing deep thaw and thus, attenuating reductions in soil moisture.

4.3 Effects of snow dynamics on active layer depth, soil climate and soil carbon

Increased duration of the snow-free season associated with future warming had a profound impact on ALD, soil climate and soil OC dynamics, although changes in snow depth alone had minimal effects. The longer snow-free season resulted in greater warming of soil, thawing of the near-surface permafrost, and soil OC losses relative to present-day conditions. However, our model simulations did not consider feedbacks associated with changes to snow seasonality, such as mid-summer drought (Welp et al., 2007) and increases in wildfire frequency (Kasischke and Turetsky, 2006). To more accurately assess the net effect of snow on soil OC dynamics in the boreal region, models are needed that consider the combined effects of snow, disturbance and hydrology (e.g., Yi et al., 2010).

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Variation in snow dynamics across boreal successional sequences has received little attention in the literature (see Euskirchen et al., 2009). Field measurements in this study show that winter soil temperatures were consistently warmer at the 2003 burn relative to the unburned mature stand and the 1967 burn (Supplement Fig. 1), despite similar snow depths and seasonal dynamics across sites (Fig. 3a). These observations indicate that the magnitude of insulation is affected by the thermal properties of not only the snow, but also the underlying soil layers (Lachenbruch, 1959). This phenomenon is characterized by the parameter μ (Sasonova and Romanovsky, 2003), which quantifies the difference in thermal properties (heat capacity, thermal conductivity) between snow and the underlying soil layer. Higher μ values reflect increased insulation, and as a result, the thermal effect of snow should be more pronounced in recently burned stands (with thin organic horizons) than in mature stands (with thick organic horizons). These observations highlight the overriding importance of organic horizon thickness (Fig. 5), and thus stand age and fire severity, as important controls on soil thermal dynamics during winter and summer.

4.4 Comparing the effects of active layer depth and soil climate on soil carbon loss

The combined effect of ALD and soil climate on soil OC was generally less than the sum of each individual effect (ALD effect+soil climate effect; Figs. 8 and 9), suggesting that these factors are not distinct but have overlapping properties. While many studies in the boreal region have documented the soil climate effect on decomposition (Dioumaeva et al., 2003; Wickland and Neff, 2007; O'Donnell et al., 2009b; Karhu et al., 2010), the nature and importance of the ALD effect is less clear. We suggest that the ALD effect integrates three factors that govern soil OC cycling in the northern permafrost region. First, the ALD effect reflects the distribution of OC into unfrozen and frozen soil, which in turn, determines the size of the OC pool (C) and contributes to the decomposition rate ($-kC$; O'Donnell et al., 2010). Second, the ALD effect incorporates the impact of soil phase (i.e., frozen vs. unfrozen) on the temperature sensitivity of decomposition

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rates (see Rivkina et al., 2000; Mikan et al., 2002). Third, the ALD effect also reflects differences in the inherent decomposability of organic matter substrates in unfrozen ($k_{\text{active-layer}}$) and frozen mineral soil ($k_{\text{permafrost}}$; Waldrop et al., 2010). While the first property of the ALD effect is distinct from the soil climate effect, the other two properties overlap with the soil climate effect. Inter-annual variability in ALD likely drives small changes in the distribution OC between unfrozen and frozen pools, whereas periodic disturbances (e.g., wildfire, erosion) may result in large transfers of OC from permafrost to and from the active layer.

5 Conclusions

Recent field studies and modeling efforts have helped shape our understanding of soil OC dynamics in relation to wildfire and permafrost dynamics in the boreal region (Carrasco et al., 2006; Harden et al., 2006; Fan et al., 2008; Yi et al., 2009a, 2010; O'Donnell et al., 2010). Our findings illustrate that atmospheric warming and increased fire severity will likely drive larger OC losses from boreal soils than from changes in snow depth or soil moisture. Moreover, our findings explore the complex interactions among climate and disturbance factors, and in turn, draw attention to how these interactions function to increase soil OC loss. For instance, future increases in fire severity associated with a warmer and drier climate resulted in large OC losses through the deep combustion of organic soil horizons and subsequent warming and thawing of permafrost soils. Future decreases in fire severity associated with a warmer and wetter climate also resulted in similar OC losses, both by alleviating the moisture limitation of decomposition and by increasing heat conduction through wet organic soils. In each case, OC loss was governed by interannual changes in active layer depth, which determines the OC pool size in frozen and unfrozen mineral soil, and by the response of soil microbes to changing soil climate. Together, these findings highlight the need for modeling frameworks that consider the combined effects of fire, soil hydrology, permafrost and snow dynamics (e.g., Yi et al., 2010).

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Table 1. Climate and disturbance scenarios tested using GIPL and Fire-C models.

Climate/disturbance scenarios	Description	Source data
Air temperature		
A2 scenario	“Business as usual” scenario following IPCC (Christensen et al., 2007)	Fig. 2
B1 scenario	Reduced greenhouse gas emission following IPCC (Christensen et al., 2007)	Fig. 2
Soil moisture		
Wetter	25% increase in VWC of organic soil horizons relative to present-day values	Supplement Fig. 3
Drier	25% decrease in VWC of organic soil horizons relative to present-day values	Supplement Fig. 3
Snow dynamics		
Deeper snowpack	25% increase in snow depth relative to present-day values	Fig. 3
Shallower snowpack	25% decrease in snow depth relative to present-day values	Fig. 3
Fire severity		
Lower severity	25% of organic horizon combusted during fire	Harden et al. (2000)
Higher severity	77% of organic horizon combusted during fire	Kane et al. (2007)
Multiple factors		
High severity×drying×warming	Interactive effects of high severity fires (77%), warming (A2), and drying	
Low severity×wetting×warming	Interactive effects of low severity fires (25%), warming (A2), and wetting	
Longer snow-free season	Snowmelt 21 days earlier, snow return 20 days later than present day (Euskirchen et al., 2009)	Fig. 3

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Table 2. Organic horizon thicknesses and active layer depths across upland fire chronosequence.

Site	<i>n</i>	Live/dead moss (cm)	Fibric (cm)	Mesic/humic (cm)	Organic horizon (cm)	Active layer depth (cm)
Unburned mature	33	6 ± 2	8 ± 4	10 ± 6	24 ± 5	45 ± 8
2003 burn	23	3 ± 4	5 ± 4	6 ± 3	14 ± 5	66 ± 12
1990 burn	10	4 ± 3	4 ± 5	2 ± 3	9 ± 4	77 ± 19
1967 burn	11	4 ± 2	4 ± 3	6 ± 2	15 ± 4	54 ± 7

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Table 3. GIPL model parameterizations at three study sites across fire chronosequence at Hess Creek, AK.

Site	Horizon thickness ^a (cm)	Volumetric water content ^b (%)	Unfrozen thermal conductivity ^c (W m ⁻¹ K ⁻¹)	Frozen thermal conductivity ^d (W m ⁻¹ K ⁻¹)
Unburned mature				
Live/dead moss	5	20	0.07	0.15
Fibric	11	25	0.15	0.20
Mesic/humic	8	30	1.40	2.00
2003 burn				
Fibric	8	49	0.26	0.75
Mesic/humic	5	55	0.25	0.75
1967 burn				
Live/dead moss	6	20	0.25	1.20
Fibric	5	25	0.25	1.50
Mesic/humic	9	68	0.40	1.40

Note: Parameterizations for specific heat capacity and unfrozen water content for each horizon were taken from calibration conducted by Romanovsky and Osterkamp (2000).

^a Horizon thickness reflect average measurements for each horizon type as measured across the Hess Creek chronosequence (O'Donnell et al., 2010).

^b VWC values vary with one standard deviation of the summer mean VWC, as measured in this study (Supplement Fig. 3). VWC data calibrated following O'Donnell et al. (2009a).

^c Unfrozen thermal conductivity values were calculated from the VWC measurements following the equations by O'Donnell et al. (2009b).

^d Frozen thermal conductivity values were prescribed following Romanovsky and Osterkamp (2000).

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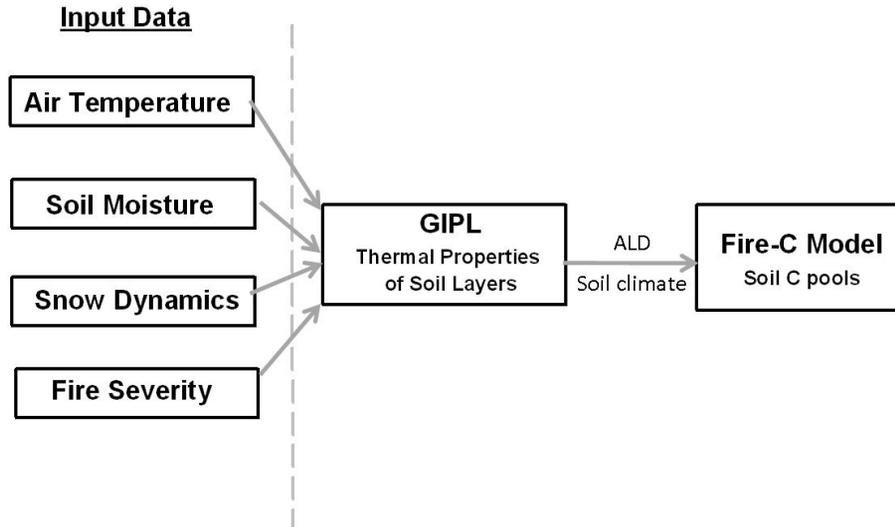


Fig. 1. The modeling framework of this study in which Geophysical Institute Permafrost Laboratory (GIPL) model was calibrated from measured temperature input data. GIPL was then used to simulate active layer depth (ALD) and soil temperature profiles over a variety of climate and soil conditions. The Fire-C model used GIPL results to simulate changes in soil OC storage over multiple fire cycles based on ALD and soil temperature (August mean monthly temperature). Input data include air temperature, soil moisture, snow depth and fire severity (carbon fraction combusted during a fire event) from the Hess Creek study area collected from 2007–2009. Sensitivity analyses were conducted (see Table 1 for scenario details) using the GIPL model to assess the impact of input variables on ALD and soil temperatures. ALD and soil temperature data were then used to drive soil C turnover in the Fire-C model.

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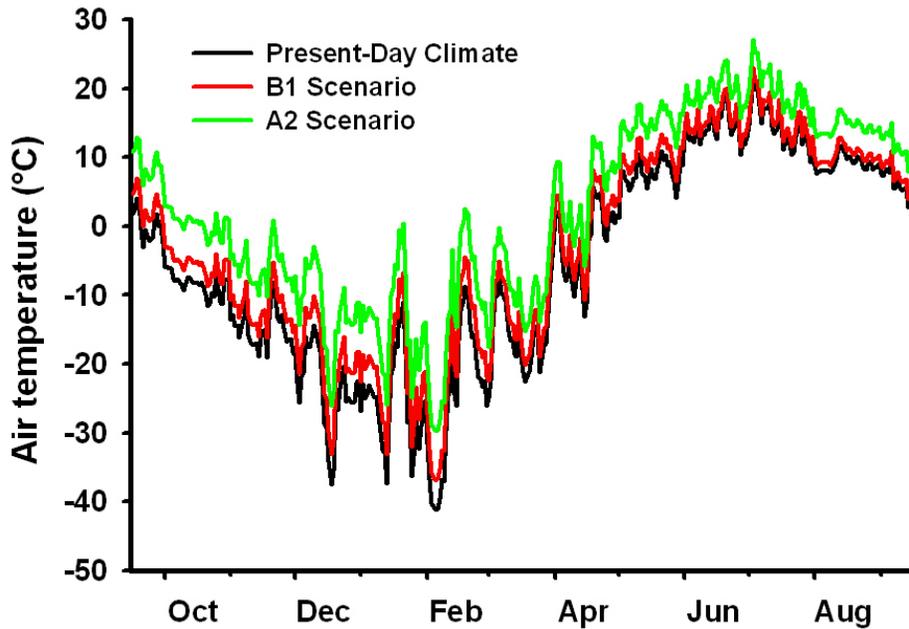


Fig. 2. Future air temperature scenarios for the year 2100 relative to present-day climate at Hess Creek study region. The B1 scenario reflects future air temperatures under reduced carbon emissions, whereas the A2 scenario reflects carbon emissions following “business as usual” trends.

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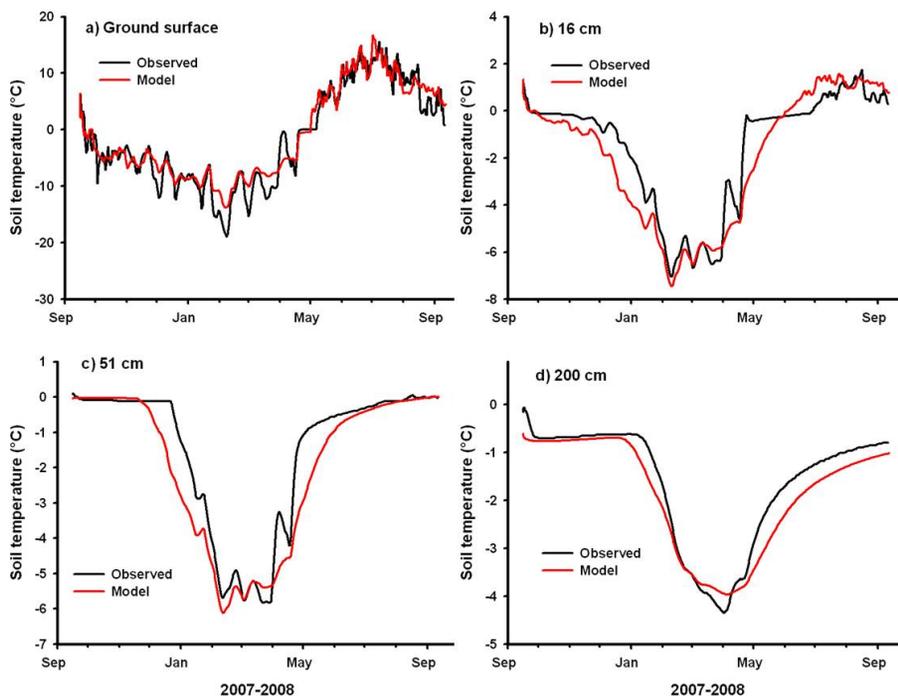


Fig. 4. Comparison of model soil temperature outputs (from GIPL2) and observed soil temperatures at different depths at the unburned mature site near Hess Creek. Model was separately parameterized for 2003 burn and 1967 burn soil profiles.

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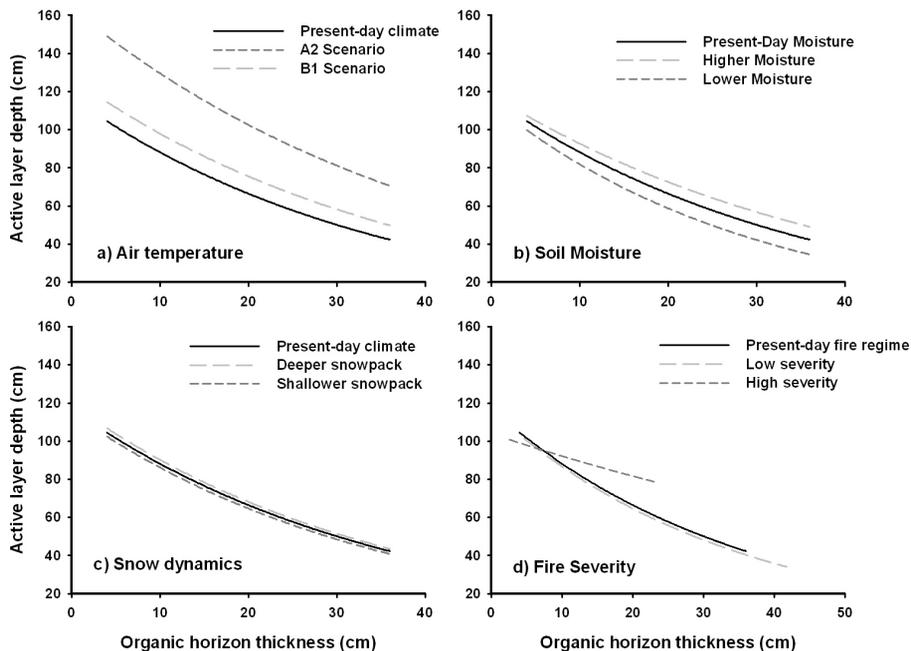


Fig. 5. Sensitivity of the active layer depth (ALD) to temperature, soil moisture, snow dynamics and fire severity. The relationship between ALD and organic horizon thickness was modeled using GIPL across a range of future air temperatures (**a**; IPCC scenarios), moisture conditions (**b**), snow (**c**), and fire severity (**d**). Output data from each scenario were then fit with the exponential equation: $ALD = a \times e^{-b \times OHT}$ and are summarized in Table 4. Equations were then used to run Fire-C model to evaluate sensitivity of soil carbon to various climate scenarios.

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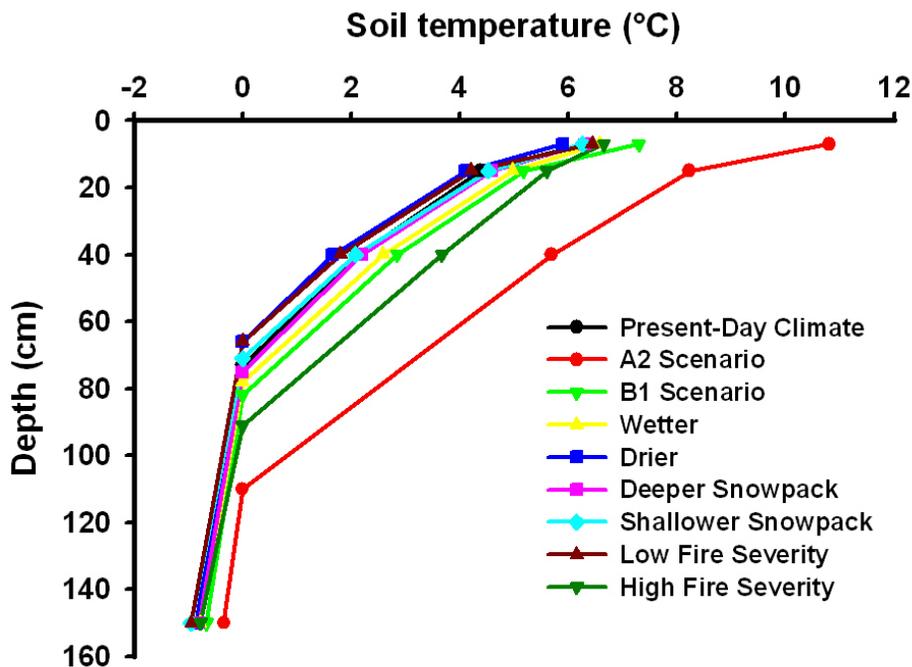


Fig. 6. Sensitivity of mean monthly temperature (MMT) in August to a suite of climatic and fire scenarios as modeled from GIPL. The depth of temperature data reflect positioning in shallow organic horizons (<10 cm), deep organic horizons (15 cm), unfrozen mineral soil in the active layer (40 cm), and permafrost (1.5 m). August MMT were used at each depth were used to drive the temperature sensitivity of organic matter decomposition (Eqs. 1 and 2) in the Fire-C model across this suite of climate and disturbance scenarios.

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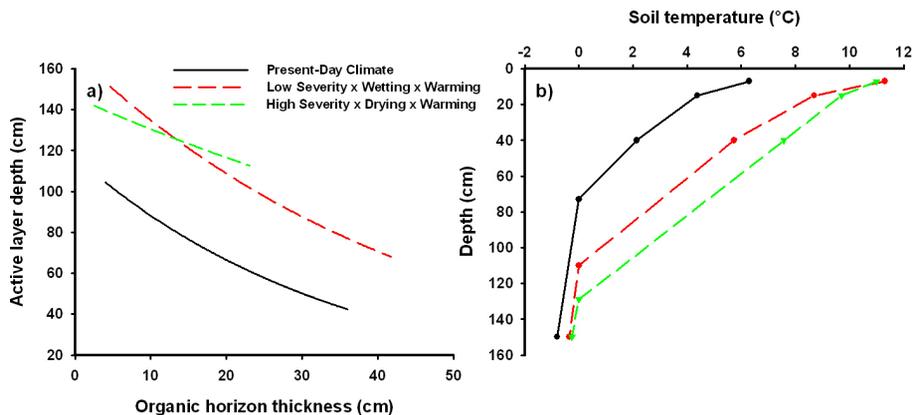


Fig. 7. Sensitivity of ALD (a) and mean monthly soil temperature in August (b) to multiple interacting factors. Using the GIPL model, we tested the effects of low fire severity, wetter soil conditions and warmer air temperatures (A2 scenario), high fire severity, drier soil conditions and warmer air temperatures, and changes in snow seasonality (following Euskirchen et al., 2009) associated with future warming (A2 scenario).

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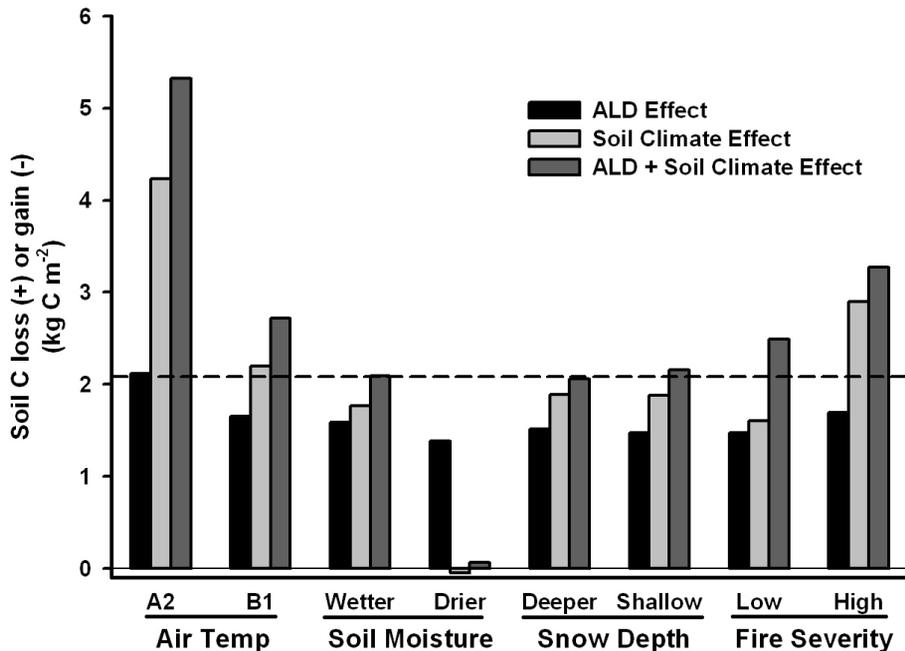


Fig. 8. The effect of single climate factors (air temperature, soil moisture, and snow depth) or fire severity on soil C storage. Soil C loss or gain was calculated as the difference in total soil C stocks in top 2 meters of soil between model years 6500 and 7000 years (or approximately 3 fire cycles after step-change in climate/fire). To calculate the active layer depth (ALD) effect, we used only modified the relationship between ALD and organic horizon thickness (from Table 4) for each scenario in the Fire-C model. To calculate the soil climate effect, we modified decomposition constants within the Fire-C model to reflect future changes in soil temperature (Fig. 6) and/or soil moisture (Table 1). To calculate the ALD + soil climate effect, we modified both the ALD-OHT relationship and the decomposition constants within the Fire-C model. The dashed green line represents total OC losses (ALD+soil climate) when present-day climate conditions (with no changes in snow, moisture or fire) were run for 3 fire cycles.

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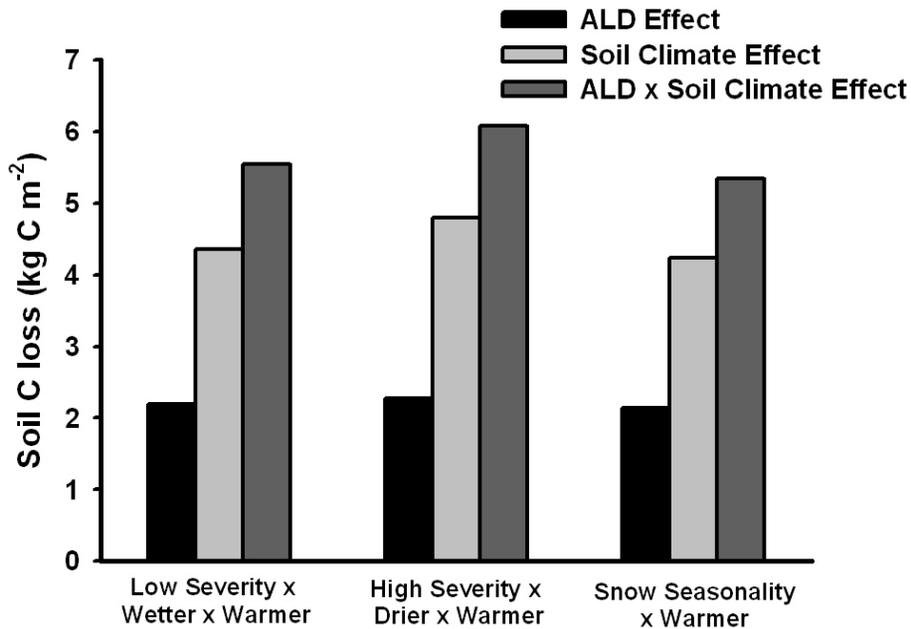


Fig. 9. The effects of combined climate/fire factors on soil C storage. The three scenarios reflect reduced fire severity associated with warmer and wetter soil conditions, higher fire severity associated with warmer and drier conditions, and longer snow-free season associated with warmer air temperatures. The ALD and soil climate parameterizations were derived from GIPL output as illustrated in Fig. 7.

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